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MONTHLY WEATHER REVIEW

First published in 1872, the Monthly Weather Review serves as a medium of publication for technical contributions in the field of meteorology, principally in the branches of synoptic and applied meteorology. In addition each issue contains an article descriptive of the atmospheric circulation during the month over the Northern Hemisphere with particular reference to the effect on weather in the United States. A second article deals with some noteworthy feature of the month's weather. Illustrated. Annual subscription Domestic, $3.50; Foreign, $4.50. Subscription to the *Review* does not include the *Supplements* which have been issued irregularly and are for sale separately.

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(Continued on inside back cover)

The Weather Bureau desires that the *Monthly Weather Review* serve as a medium of publication for original contributions within its field, but the publication of a contribution is not to be construed as official approval of the views expressed.

The issue for each month is published as promptly as monthly data can be assembled for preparation of the review of the weather of the month. In order to maintain the schedule with the Public Printer, no proofs will be sent to authors outside of Washington, D. C. Requests for a limited number of reprints of the more important contributions can be granted; such requests should be made by the author on the first page of his manuscript.

The printing of this publication has been approved by the Director of the Bureau of the Budget, February 11, 1952

DEPARTMENT OF COMMERCE
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WEATHER BUREAU
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VARIABILITY OF ANNUAL PRECIPITATION IN CANADA

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[Manuscript received March 9, 1953]

ABSTRACT

From computed values of the coefficient of variation of precipitation for Canada and northern United States, the distribution of this variable across Canada is determined. The minimum variability increases as one moves poleward. Also the variability is great in the central prairies. The pattern among the mountains and valleys of British Columbia is irregular.

INTRODUCTION

In the study of the rainfall of a district the two most important features are the average for the year and the normal variation through the year. The variation from one year to another is also very important but one on which little work has been done. This variation determines the types of industry that can thrive in the vicinity. If the industry is such that it needs each year the median rainfall or more for success, it will have difficulty in one year out of two. Another industry which demands only the amount given by the first quartile will have sufficient moisture three years out of four.

RESULTS OF PREVIOUS STUDIES

The only previous attempt to determine the distribution of the variation of precipitation in Canada was made by Taylor [1]. His work was based on only 30 stations. He states that in most cases the annual rainfalls for 20 years since 1900 were used.

Before proceeding with a more complete study of the variation, it was felt that an analysis of the different measures was necessary in order to determine their limitations and to select the best. This earlier study was reported by Longley [2]. As a result of this study which considered the advantages and disadvantages of different measures of variability, it was decided that the measure to be used was the coefficient of variation, v_σ, i. e., the standard deviation divided by the mean. In most previous studies, including the one by Taylor, the measure used was the relative variability, v_r, i. e., the mean deviation divided by the mean. Since approximately $v_r = \frac{4}{5} v_\sigma$ the results obtained may be compared with previous studies if it is desired.

Another result from the previous study was the decision to limit the calculation of v_σ to those stations where 35 years of data were available, or, in districts where such stations were not available, to use no station which had less than 30 years of record. For shorter records, the variation because of the error of sampling was too great to consider the values reliable.

DATA USED

After the measure and the minimum length of record to be used were determined, the value of v_σ was computed for 142 Canadian stations and 34 stations from the neighboring States of the United States. The data for the Canadian stations were obtained from the published and unpublished records of the Meteorological Service of Canada. Records for every available year to 1950 were used. The data from the United States were obtained from the "Local Climatological Summaries with Comparative Data" published under the direction of the United States Weather Bureau. Data used covered the period 1900–1950.

The data for British Columbia and Washington have already been used in the study by Longley. In that report there was an error which was brought to the author's attention by Mr. Philip Williams, U. S. Weather Bureau, Salt Lake City, Utah. The mean annual precipitation at Tatoosh Island should be 74.88 inches, not 30.64, and coefficient of variability 14.8, not 36.2. This change will also cause a change in the isolines of figure 2 of that study, and give a distribution as shown in figure 1 of this report.

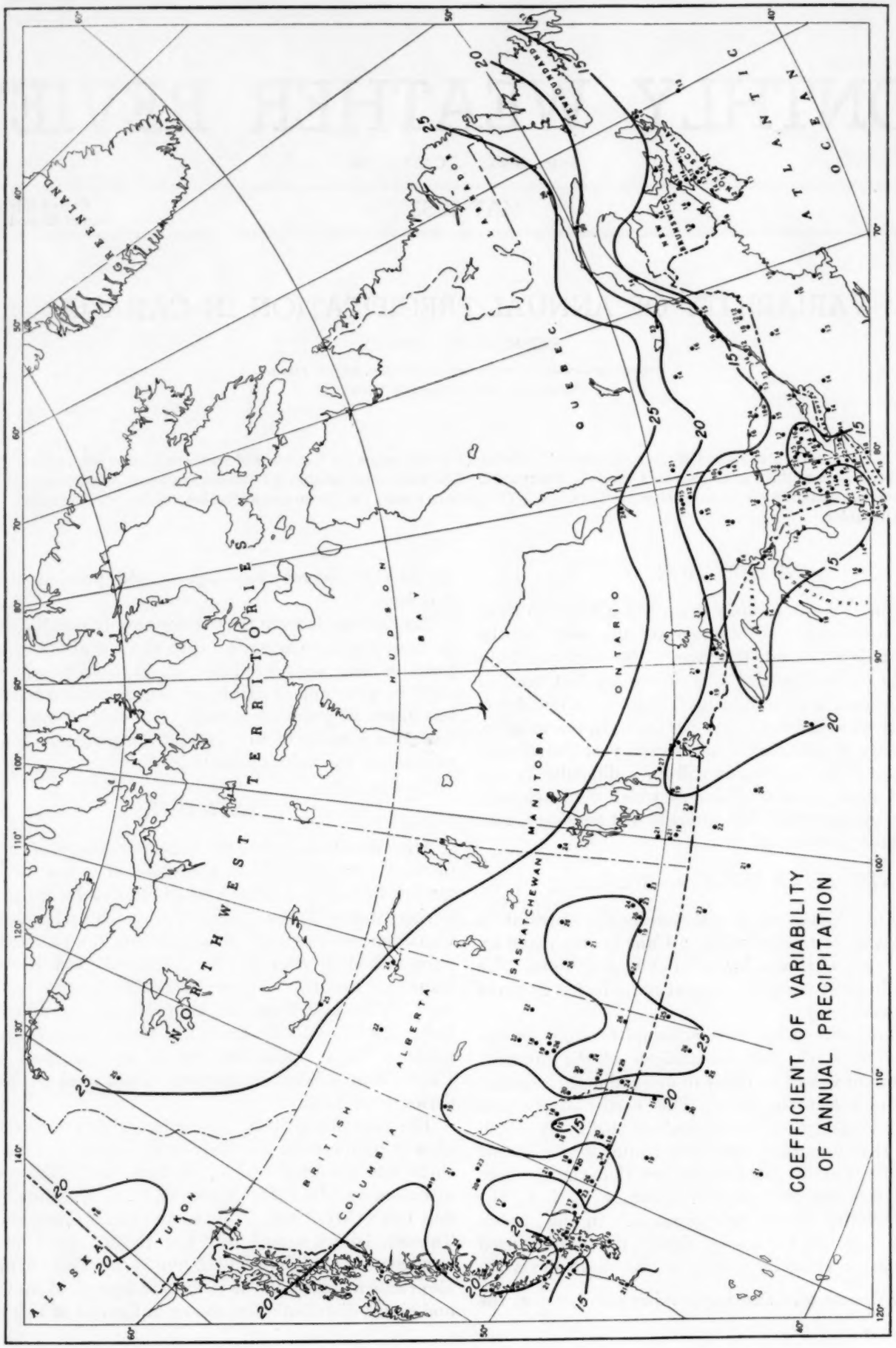

FIGURE 1.—Coefficient of variation (σ, percent) of annual precipitation in Canada.

TABLE 1.—*Mean, standard deviation, and coefficient of variation of annual precipitation in Canada and northern United States*

Station	Number of years	Mean precipitation (in.)	Standard deviation (in.)	Coefficient of variation (percent)
British Columbia				
Agassiz	57	58.68	12.70	22
Alberni	48	66.50	12.36	19
Atlin	31	10.90	2.70	24
Barkerville	56	40.21	9.37	23
Bella Coola	45	54.90	11.90	22
Big Creek	42	12.27	2.11	17
Clayoquot	41	106.3	22.24	21
Fort St. James	55	15.62	3.02	19
Garry Point	48	36.46	5.87	16
Glacier	38	56.90	8.39	15
Golden	35	18.31	3.20	18
Greenwood	35	17.04	5.77	34
Hedley	39	11.45	2.96	26
Hope	36	56.39	9.72	17
Kamloops	55	10.21	2.18	21
Kelowna	46	12.22	2.40	20
Masset	46	55.17	9.38	17
Nanaimo	49	37.91	7.75	20
Nelson	44	26.32	5.34	20
New Westminster	68	57.06	7.47	13
Prince George	33	22.31	4.61	21
Prince Rupert	36	93.88	13.54	14
Princeton	53	13.17	3.01	23
Quatsino	46	92.74	18.55	20
Quesnel	46	16.88	3.46	20
Revelstoke	32	39.43	6.44	16
Rossland	38	29.08	5.32	18
Vancouver	42	58.02	8.12	14
Vernon	48	14.80	2.70	18
Victoria	52	26.90	5.09	19
Yukon				
Dawson	49	12.81	2.12	16
Northwest Territories				
Fort Good Hope	42	11.62	2.93	25
Hay River	37	11.68	2.93	25
Alberta				
Alix	38	16.15	3.62	22
Banff	55	18.86	3.80	20
Beaverlodge	30	17.29	3.33	19
Calgary	66	16.70	4.98	30
Edmonton	68	17.46	3.79	22
Fort Chipewyan	30	11.85	2.80	24
Fort Vermilion	39	12.05	2.71	22
Gleichen	37	14.72	3.48	24
Hillsdown	43	17.22	4.48	26
Lacombe	41	17.74	3.45	19
Macleod	39	14.89	4.38	29
Medicine Hat	67	12.91	3.40	26
Pekisko	39	24.17	6.28	26
Pincher Creek	40	20.35	4.90	24
Ranfurly	46	17.70	3.92	22
Wetaskawin	37	17.43	3.12	18
Saskatchewan				
Battleford	60	13.35	3.52	26
Grenfell	44	17.79	4.05	23
Indian Head	53	17.28	3.90	23
Moose Jaw	42	14.82	2.78	19
Prince Albert	66	15.73	4.09	26
Qu'Appelle	67	18.32	4.39	24
Regina	62	14.37	3.80	26
Saskatoon	44	14.43	3.08	21
Swift Current	65	15.04	3.56	24
Manitoba				
Brandon	59	18.11	3.73	21
Minnedosa	68	17.29	3.62	21
Morden	64	19.08	4.35	23
Pinawa	33	15.18	4.05	27
The Pas	38	16.71	3.98	24
Treesbank	51	18.29	3.52	19
Winnipeg	79	20.83	3.91	19
Ontario				
Agincourt	53	30.44	4.27	14
Alton	49	32.89	3.46	10
Barrie	44	31.70	5.69	18
Beatrice	74	40.88	5.61	14
Biscotasing	38	28.66	5.11	18
Brantford	47	30.54	4.83	16
Brucefield	44	33.16	5.31	16
Chatham	59	29.28	4.93	17
Clontarf	57	30.80	4.67	15
Cochrane	33	30.56	5.75	19
Coldwater	37	34.69	7.41	21
Fort Francis	33	27.05	5.39	20
Goderich	71	30.54	5.93	19
Guelph	59	29.89	5.03	17
Haileybury	46	31.40	3.22	10

TABLE 1.—*Mean, standard deviation, and coefficient of variation of annual precipitation in Canada and northern United States*—Con.

Station	Number of years	Mean precipitation (in.)	Standard deviation (in.)	Coefficient of variation (percent)
Ontario—Continued				
Haliburton	35	33.33	4.28	13
Iroquois Falls	34	33.03	4.03	12
Kakabeka Falls	41	26.92	4.35	16
Kenora	44	25.23	4.72	19
Kingston	77	33.10	4.74	14
Lakefield	47	28.71	4.71	16
Lindsay	68	33.13	5.84	18
London	66	38.24	5.30	14
Lucknow	63	36.90	4.88	13
Mine Centre	33	27.26	5.19	19
Montreal River	40	32.20	4.58	14
Moosonee	31	26.50	6.34	24
Morrisburg	35	38.93	5.00	13
Orillia	57	32.80	4.97	15
Ottawa	77	34.12	4.42	13
Owen Sound	56	36.09	6.45	18
Paris	61	34.66	4.18	12
Parry Sound	76	38.63	4.84	12
Pelee Island	36	29.94	5.41	18
Pembroke	51	31.83	7.48	24
Peterborough	75	30.77	4.51	15
Port Arthur	73	24.91	4.88	20
Port Dover	64	33.16	6.14	18
Port Stanley	51	35.13	4.48	13
Renfrew	57	27.76	4.96	18
Ridgetown	41	33.33	4.50	14
Rutherglen	35	30.56	4.69	15
Savanne	40	26.05	5.39	21
Schreiber	38	29.74	6.83	23
Simcoe	48	34.73	7.67	22
Southampton	77	35.24	4.69	13
Stonecliff	36	28.34	4.64	16
Stratford	83	38.07	5.18	14
Tobermory	32	32.50	4.46	14
Toronto	106	32.20	5.05	16
Turbine	34	32.06	3.92	12
Wallaceburg	32	28.56	6.20	22
Wawaitin Falls	33	30.86	4.66	15
Welland	60	34.12	5.26	15
White River	60	28.30	5.32	19
Windsor	53	31.71	4.37	14
Woodstock	76	32.68	4.81	15
Quebec				
Abitibi	38	26.82	5.57	21
Anticosti	66	31.23	6.29	20
Brome	58	39.34	7.34	18
Chicoutimi	50	30.66	6.94	23
Father Point	74	33.50	4.84	14
Gouin Dam	35	38.27	5.22	14
La Tuque	35	33.56	5.51	16
Lennoxville	35	40.07	4.90	12
Montreal	79	41.07	4.96	12
Natashquan	29	36.12	9.02	26
Quebec	80	37.76	5.71	15
Ste. Anne de Bellevue	33	38.17	5.18	14
Ste. Anne de la Pocatiere	35	37.30	5.80	16
Shawinigan Falls	41	36.30	5.76	16
Sherbrooke	47	38.50	4.74	12
New Brunswick				
Chatham	77	39.86	5.80	14
Fredericton	74	42.08	5.63	13
Moncton	50	36.96	5.27	14
Saint John	79	46.59	6.06	13
Sussex	50	39.42	5.46	14
Nova Scotia				
Annapolis Royal	33	41.41	6.01	14
Halifax	77	55.61	6.31	11
Kentville	37	40.82	5.56	13
Liverpool	37	57.72	6.33	11
Nappan	37	38.23	4.54	12
Parrsboro	41	42.84	6.19	14
Sable Island	53	50.16	6.59	13
Sydney	78	50.72	6.83	14
Truro	56	41.36	5.91	14
Windsor	37	39.80	5.58	14
Wolfville	45	39.97	5.33	13
Yarmouth	70	51.50	6.00	12
Prince Edward Island				
Charlottetown	75	40.27	6.48	16
Newfoundland				
St. John's	65	53.96	7.08	13
United States				
Albany, N. Y.	51	33.17	4.31	13
Alpena, Mich.	51	26.66	3.18	12
Bismarck, N. Dak.	51	15.71	3.32	21

TABLE 1.—*Mean, standard deviation, and coefficient of variation of annual precipitation in Canada and northern United States*—Con.

Station	Number of years	Mean precipitation (in.)	Standard deviation (in.)	Coefficient of variation (percent)
United States—Continued				
Boise, Idaho	51	12.17	2.50	21
Boston, Mass.	51	38.54	5.11	13
Burlington, Vt.	51	31.87	4.55	14
Buffalo, N. Y.	51	33.26	4.71	14
Cleveland, Ohio	51	32.76	5.73	18
Concord, N. H.	51	36.42	5.93	16
Detroit, Mich.	51	31.44	4.67	15
Devil's Lake, N. Dak.	46	17.33	3.84	22
Duluth, Minn.	51	26.67	3.89	15
Eastport, Maine	51	36.22	6.42	18
Erie, Pa.	51	35.66	6.42	18
Fargo, N. Dak.	51	20.78	5.34	26
Grand Rapids, Mich.	51	32.19	5.26	16
Great Falls, Mont.	51	14.97	3.93	26
Green Bay, Wis.	51	26.89	4.70	18
Havre, Mont.	51	12.65	3.19	25
Helena, Mont.	51	12.29	3.13	25
Kalispell, Mont.	52	14.86	3.16	21
Lansing, Mich.	51	31.36	4.40	14
Marquette, Mich.	51	31.81	4.44	14
Minneapolis, Minn.	51	26.79	5.16	19
Missoula, Mont.	50	13.76	2.73	20
North Head, Wash.	51	54.38	11.18	21
Portland, Maine	51	41.52	6.39	15
Rochester, N. Y.	51	31.87	4.30	14
Sault Ste. Marie, Mich.	51	29.26	3.67	12
Seattle, Wash.	51	32.15	6.21	19
Spokane, Wash.	51	15.05	2.87	19
Syracuse, N. Y.	48	35.54	4.89	14
Tatoosh Island, Wash.	48	74.88	11.08	14
Williston, N. Dak.	51	14.36	3.29	23

Unfortunately, the 142 Canadian stations were not distributed uniformly. Over one-third were in Ontario, and there were only two in the Northwest Territories. Also there were only two in Ontario and Quebec north of the 50° parallel. This distribution reflects the interest in the weather early in the century, for a station must have been established before 1920 to be included in this study.

The values for v_s for the different stations are given to the nearest percent in table 1 and are plotted on figure 1. Isolines are drawn for as much of Canada as it was thought the data justified.

ANALYSIS OF RESULTS

One of the first conclusions in examining the results is that the data are very irregular. From the meteorological point of view, the variation for southern Ontario should not be great, for the same storms sweep over all stations. The adjacent lakes do cause small differences in rainfall, but most of the area has an annual precipitation between 30 and 35 inches. Yet in a line from Simcoe on Lake Erie to Coldwater on Georgian Bay, there is the following succession of values: 18, 22, 14, 12, 17, 10, 18, 15, and 21. Again in the Minnesota-North Dakota region, Fargo, N.Dak., with a value of 26 for v_s is surrounded by values of 19, 15, 22, and 21, with the data taken for the same period of 51 years.

One cause for irregularity is inconsistent and, at times, careless observing. The data from Sable Island give one example. $v_s=13.5$ for a period of 53 years. The results for two years were discarded since the results were considered unreliable because of a leaky rain gage. Had these been included the value of v_s would have been 18.8, five percentage points greater.

There must be other courses, at present obscure, for the change in v_s from place to place. Lucknow, Goderich, and Southampton, Ontario, are located within a small radius and all have records over 60 years in length. Yet the values of v_s are respectively 13.2, 19.4, and 13.3. It is possible that in British Columbia the movement of a station from one location to another might influence the result. But in southern Ontario, where the precipitation is so uniform, a transfer over a short distance should not cause a serious break in the record.

In spite of the irregular spatial variation in v_s noted above, there is found a broad over-all picture of the change across Canada. The region where the variation is least is found in the Maritime Provinces of eastern Canada. There v_s is about ⅛. In other words, the total precipitation for the year is found two out of three times between ⅞M and 1⅛M, where M is the mean. This assumes that the distribution is approximately normal. The isoline for 15 percent extends from this district westward to include much of tha area about the Great Lakes. This district from the Great Lakes eastward is in the track of the major storms of the continent. The Mississippi Valley storms usually hit the whole area. The storms along the east coast of the United States pass over the Maritime Provinces, bringing frequent rains.

Although the British Columbia coast gets more precipitation than eastern Canada, the variability is greater. At a few stations it is below 15 percent but most of the coastal regions have a variability between 15 and 20 percent.

There are two areas where the value of v_s is over 25 percent. That means approximately that in two out of three years the annual rainfall lies between $\frac{3}{4}$ M and $\frac{5}{4}$ M. One of these areas lies along the Mackenzie River. On the map the 25 percent isoline is drawn across the map from Belle Isle, Newfoundland, to near Moosonee, and then across Ontario and Manitoba to Great Slave Lake. On the basis of the available evidence this may be right. We need further data to confirm this. The second district extends along the South Saskatchewan, and into central Montana. A further investigation is necessary to explain why this area has a greater variability of precipitation than the areas on all sides.

ACKNOWLEDGMENTS

The author wishes to acknowledge his indebtedness to Mr. C. C. Boughner and his staff of the Climatological Section for their interest and cooperation in this study, and to Mr. Andrew Thomson, Controller, Meteorological Services of Canada, for permission to publish the report.

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1. G. Taylor, *Canada*, Methuen and Co., London, 1947, pp. 69–70.
2. R. W. Longley, "Measures of the Variability of Precipitation," *Monthly Weather Review*, vol. 80, No. 7, July 1952, pp. 111–117.

THE WEATHER AND CIRCULATION OF MAY 1953[1]—

One of the Worst Tornado Months on Record

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SEVERE LOCAL STORMS

In many sections of the United States the weather of May 1953 was notable for severe thunderstorm activity and accompanying phenomena: torrential rainfall, local floods, huge hailstones, destructive winds, and devastating tornadoes. Most spectacular were the destructive tornadoes which occurred mainly across the South from Texas to Alabama, in the Plains States, and along the northern border from Montana to the Great Lakes region (fig. 1). Tornadoes normally are most frequent in May in the United States, but the total number of tornadoes reported during May 1953 was 113, the greatest number for any month of the entire period of record commencing in 1916 [1]. It must be pointed out though that strict comparison with tornado counts for earlier years may not be justified since in recent years increased population distribution and more extensive tornado reporting have probably resulted in an increase in the proportion of

[1] See Charts I–XV following p. 154 for analyzed climatological data for the month.

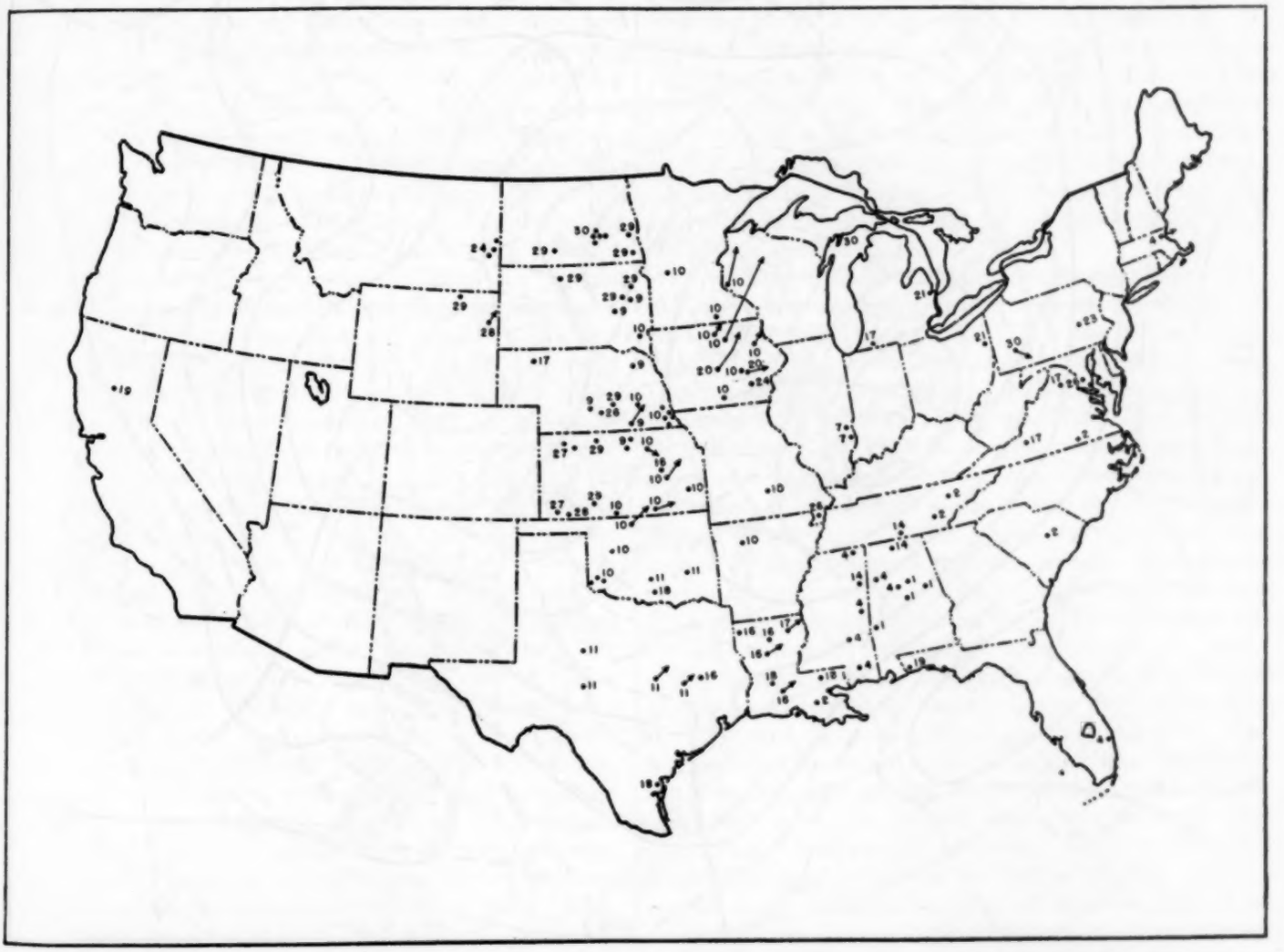

FIGURE 1.—Location of tornadoes reported over the United States during May 1953. Numbers indicate date of occurrence and arrows indicate well-marked tracks. Note the concentrations in the South Central States and northward through the Plains. Twenty-seven of the total of 113 tornadoes shown occurred on May 10. Figure is based on information provided by Climatological Information Section of Weather Bureau. These reports are not complete and it is probable that additional tornadoes occurred.

tornadoes reported. An alternate way of measuring frequency of tornadic activity is in terms of the number of days with tornadoes in the United States as a whole. For May 1953 there were 21 tornado days, a figure exceeded only by May 1933 with 23 and May 1950 with 22, and equalled in May 1951. Thus, it must be concluded that this May was one of the worst tornado months on record. On May 10 alone some 27 tornadoes were reported over the region from Oklahoma and Arkansas northward into Minnesota and Wisconsin (fig. 1). These occurred to the east and southeast of a deep cyclone which moved northward from western Kansas to eastern North Dakota (Chart X). Associated with this Low was a deep trough aloft and an intense cold front oriented North-South along which cold, dry Pacific polar air was displacing a broad southerly current of very moist maritime tropical air. On May 11, as this same cold front became stationary over central Texas, additional tornadoes developed in Texas and Oklahoma. Two of these ripped through the cities of San Angelo and Waco, Tex. The tornado in Waco resulted in 114 deaths and caused property damage estimated at about 14 million dollars—the heaviest financial setback to a city in tornado history.

Heavy rainfall for several days in central and western Louisiana and eastern Texas caused major floods in the

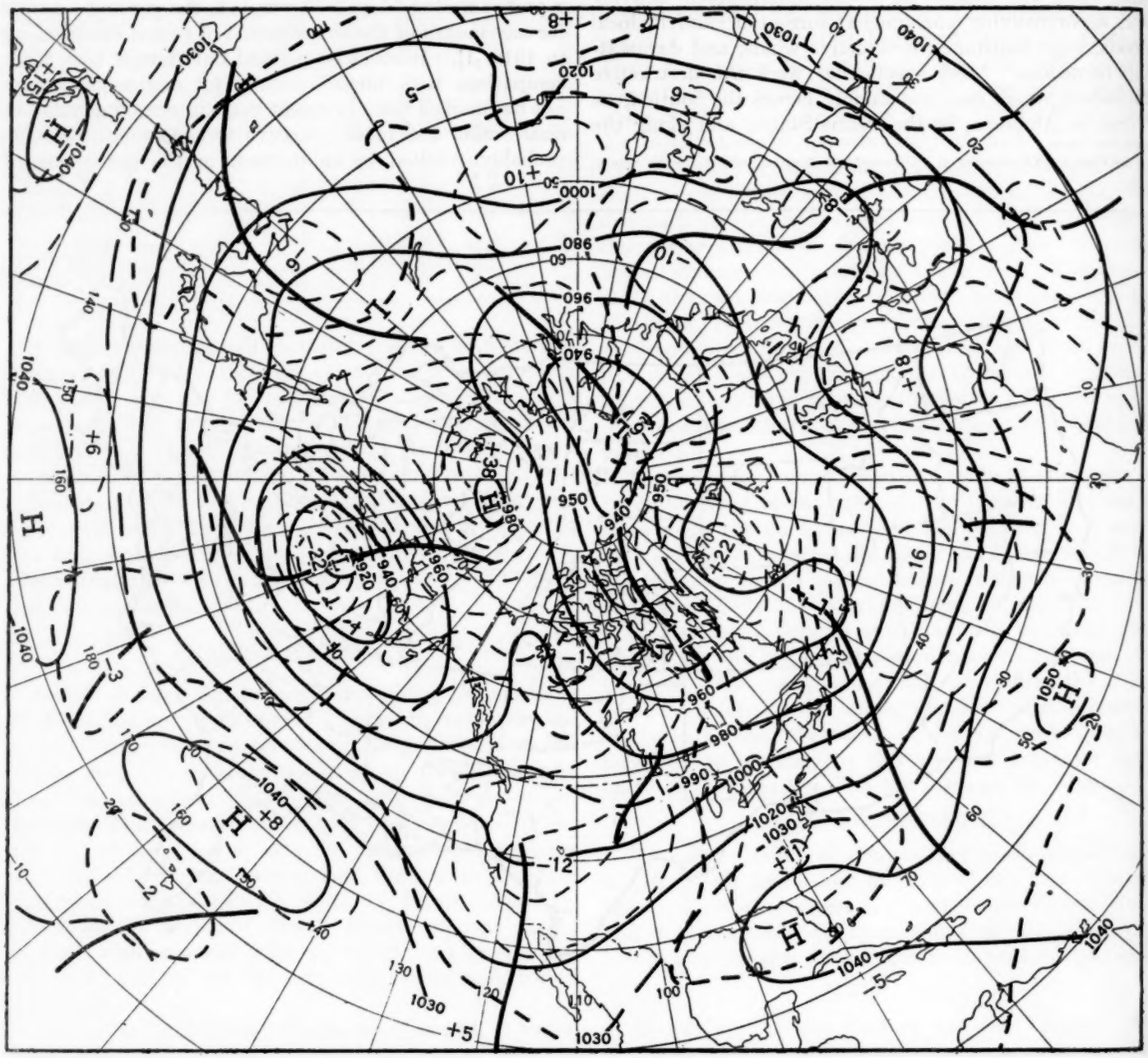

FIGURE 2.—Mean 700-mb. height contours and departures from normal (both labeled in tens of feet) for May 2–31, 1953. Basic wave pattern over United States consisted of trough in West and ridge in East with anomalous southerly flow components over middle of country. This is recognized pattern associated with severe local storminess in spring over large portions of United States.

second half of the month. Some of the rivers in this area reached their highest flood stages in 40 years. Near the end of the month heavy rains in Montana and the Northern Plains caused local flooding of rivers resulting in considerable property damage. Damaging hailstorms were in the news in May, too. On May 6, 7, and 9 hail occurred in sections of the central Atlantic Coast. Near the end of the month hailstones as big as baseballs fell in such cities as Washington and Detroit, shattering windows and denting many automobiles.

RELATED MONTHLY MEAN CIRCULATION FEATURES

The large-scale features of the 700-mb. monthly mean circulation pattern over the North American region for May provide much of the explanation for this widespread severe storminess. Figure 2 shows that the circulation over the United States was dominated by a trough in the West and a ridge in the East, both of which had height departures from normal of about equal magnitude (−12 over Nevada, +11 over South Carolina). This circulation pattern provided a stronger-than-normal flow of air from the Gulf of Mexico northward through the central part of the country, and also an abnormally strong eastward flux of cool Pacific air through the Southwest trough into the region east of the Rockies. This Pacific air remained cold and unstable in mid-troposphere in its travel across the West because of the prevailing cyclonic circulation. It is believed that a large-scale circulation pattern of this type is especially favorable for the development of severe squall lines and tornadic activity since it provides for more frequent and more intense interactions between the maritime polar and maritime tropical air masses than would normally occur. It has long been recognized, of course, that intense squall lines and tornadoes are most favored synoptically where cold, dry Pacific air masses are advected rapidly aloft over moist tropical air in lower levels (see [2]). It may be recalled that an unusually large number of tornadoes also occurred during April 1953. Although the circulation patterns in April and May were different in many respects, they were alike in that the flow of Pacific air masses into the central United States was stronger and more cyclonic than normal [3].

Even though severe storminess was frequent during May, many States with a considerable number of tornadoes did not have abnormally heavy precipitation for the month as a whole (Chart III). In fact, in most of Texas, Oklahoma, Kansas, Nebraska and Iowa, where more tornadoes than normal were observed (fig. 1), precipitation amounts generally totaled well below normal. This may, of course, be associated with the fact that tornadoes occurred only during five or six days of the month in these States. However, even on those days when several tornadoes were reported in the Plains States, 24-hour precipitation amounts rarely exceeded ½ inch. It is probable that the cause of these relatively small precipitation amounts accompanying severe storminess in the Plains lies in the considerable amount of dry Pacific air aloft and prevailing downslope motion east of the Rockies associated with circulation regimes of the type shown in figure 2.

The difference in the monthly precipitation regimes over the Southwest and Northwest is quite interesting and rather typical. It may be seen from figure 2 and Chart III that heavier-than-normal precipitation generally occurred where flow was more easterly than normal to the north of the negative height anomaly center over Nevada, and lighter-than-normal rainfall occurred to the south under fast westerly flow. The physical explanation for this relationship apparently lies in the tendency for more frequent than normal occurrence during the month of upper closed Lows near the negative height anomaly center and in the typical upward motion to the north of such systems.

TRANSITION OF CIRCULATION AND PRECIPITATION REGIMES DURING MAY

Of fundamental importance in understanding the weather of May 1953 was the gradual but considerable change in large-scale circulation pattern which took place over the North American region during the month. To illustrate these transitions and their effect on precipitation three consecutive 10-day mean 700-mb. charts and the corresponding 10-day percentage of normal precipitation are presented in figures 3 and 4 respectively. In addition, to provide further insight into these precipitation patterns, counts have been made for each 10-day period of the number of days with surface fronts (of any type) located within equal-area squares (fig. 5).

In the first third of May four troughs existed at middle and lower latitudes between the west-central Pacific and the central Atlantic (fig. 3A). Over the United States two ridges, one near each coast, flanked the major trough extending from the Northern Plains southward and southwestward to Lower California. Neither the eastern ridge nor the United States trough extended very much north of the Canadian border since a broad northwesterly flow existed over eastern Canada, and the deep cyclonic vortex centered over Baffin Island was essentially connected with the central Atlantic trough. The eastern United States ridge was of rather small amplitude from Pennsylvania southward while over the Great Lakes it had a more pronounced anticyclonic development.

Precipitation over the country in these first ten days of May was well related to these circulation features (fig. 4A). A large area from east Texas northeastward to New England and eastward to the Carolinas was covered for the most part with greater than normal amounts of rainfall. This area was located under southwesterly flow aloft which transported abundant supplies of moist tropical air northward from the Gulf of Mexico. Instrumental in the precipitation of much moisture from these air masses were the southerly components of flow, usually an indicator of upward motion, and the frontal systems in the South and East. Note how the region in which

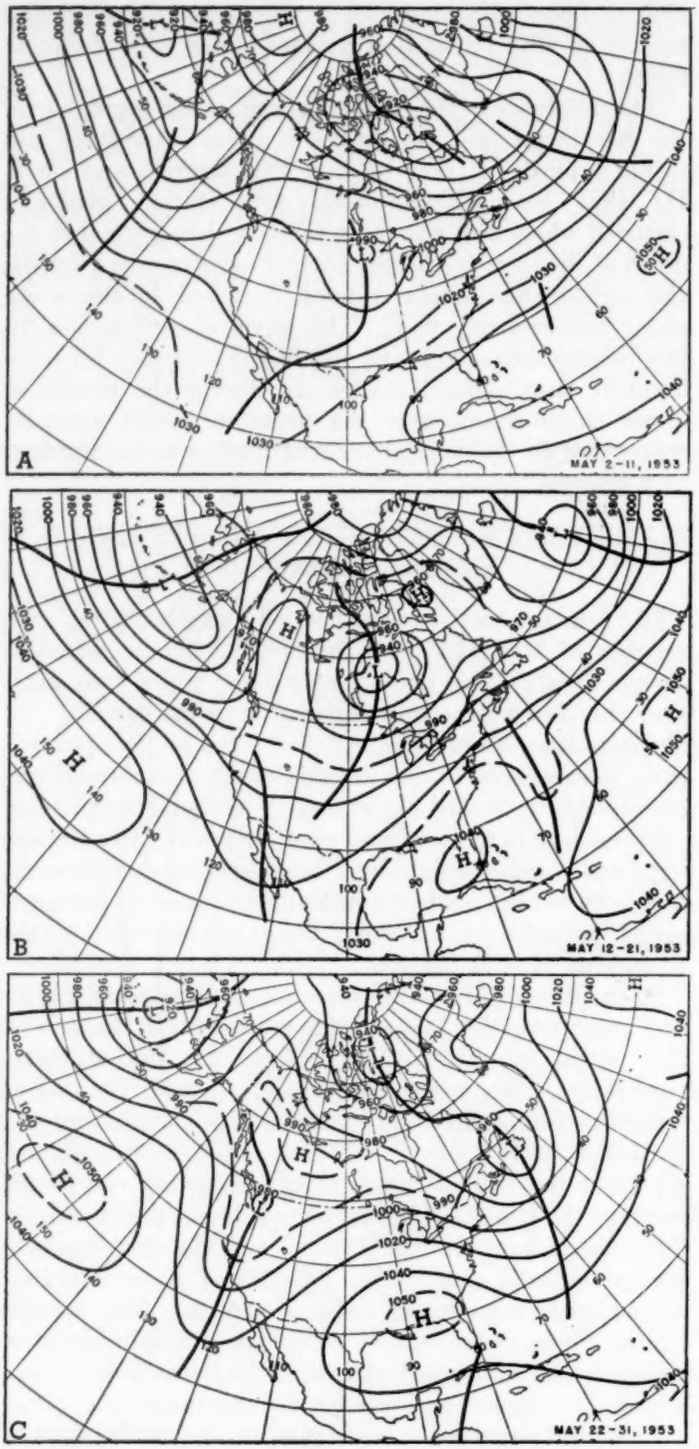

FIGURE 3.—Mean 700-mb. charts for (A) May 2-11, 1953, (B) May 12-21, 1953, and (C) May 22-31, 1953 (contours labeled in tens of feet). Note nearly complete reversal of wave pattern over United States from first to third decade of month with a large ridge replacing a trough over middle of country and a deep trough replacing a ridge near West Coast.

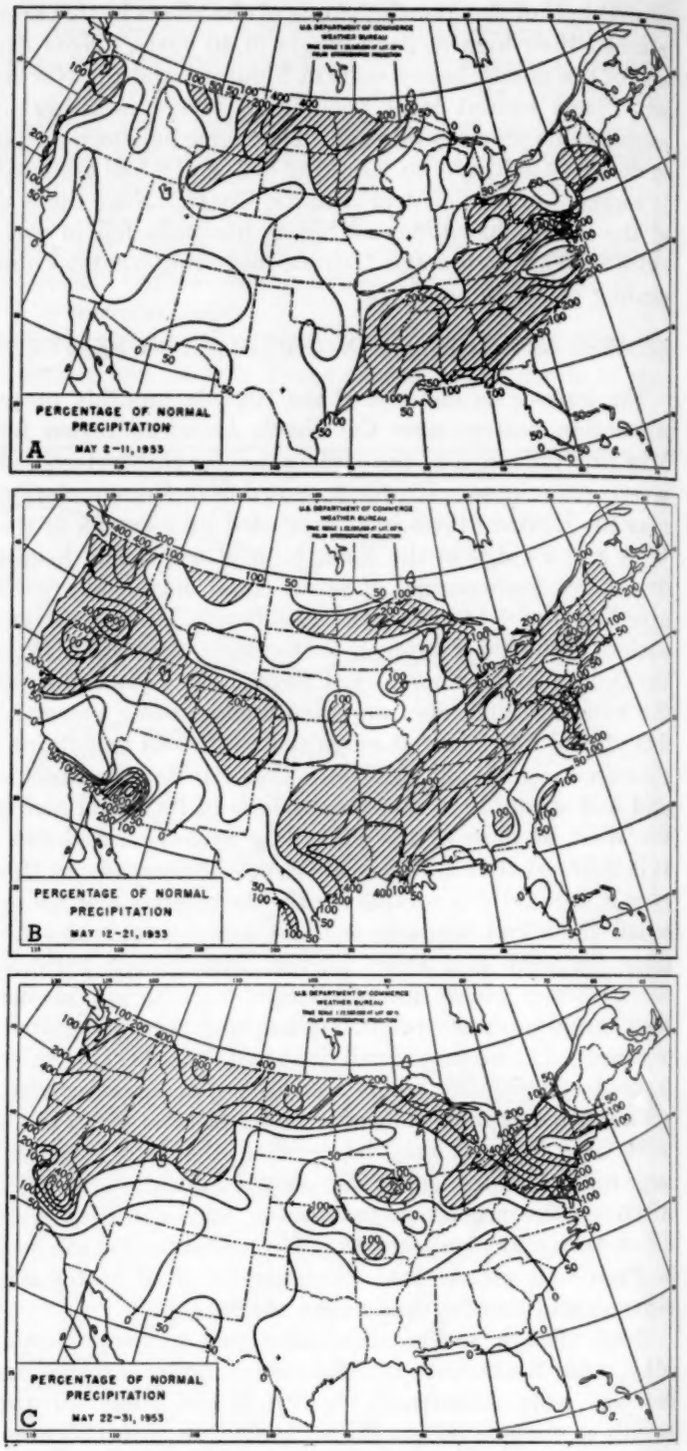

FIGURE 4.—Percentage of normal precipitation amounts for the periods (A) May 2-11 1953, (B) May 12-21, 1953, and (C) May 22-31, 1953. Regions in excess of normal are shaded. Heavy rainfall in South in first two-thirds of month was followed by complete absence of rainfall in same area during last 10 days of month. This was related to large-scale circulation changes illustrated in figure 3.

fronts were present during six or more days of the period (fig. 5A) extended eastward and northeastward from east Texas in fairly close correspondence with the area of supernormal precipitation. Over much of the Great Lakes region and parts of the Midwest precipitation was abnormally light. It is believed that the nearly separate anticyclonic circulation aloft over the Lakes provided so much subsidence that little rain fell even though there were fronts present on six or more days over the region. Over the Plains heavier-than-normal precipitation occurred mainly near the Canadian border from Montana eastward to Minnesota. This is attributable to the closed

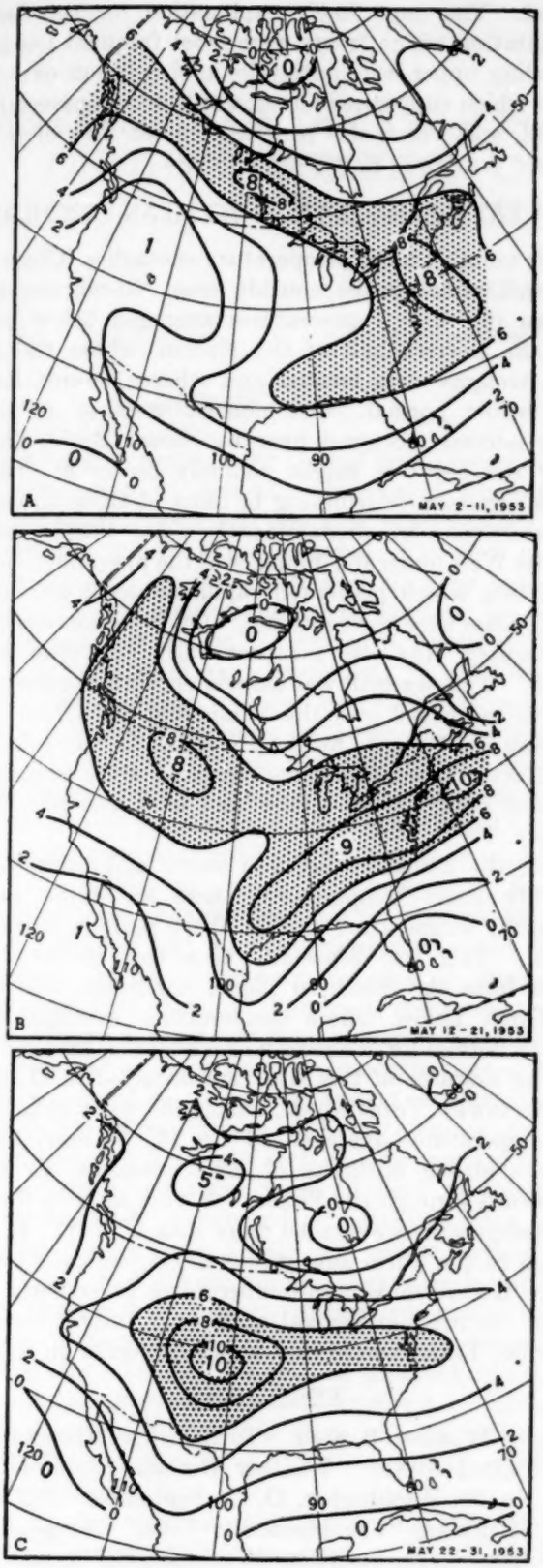

FIGURE 5.—Number of days with surface fronts (of any type) located within square areas with sides approximately 430 nautical miles in length for (A) May 2–11, 1953, (B) May 12–21, 1953, and (C) May 22–31, 1953. Frontal positions were those appearing on printed Daily Weather Maps for 1830 GMT. Areas where fronts were present on six or more days are shaded. Note great frequency of fronts over South and East in first two decades of month which was closely associated with heavy rainfall in these areas as shown in figure 4.

mean cyclonic circulation located over the North Dakota-Minnesota boundary. With such circulations moisture and precipitation are typically carried westward across the top of the Low, and it is not unusual for the greatest amounts of precipitation to be located northwest of the center, as in this case. Over most of the central and southern Plains precipitation was generally subnormal even though the trough was located over the area. As mentioned earlier many tornadoes occurred in this region during this period (fig. 1). As explained with respect to the monthly mean circulation, the dominance of dry Pacific air in this region accounts in part for both tornadic activity and subnormal rainfall. Very dry weather over most of the Rockies, the Plateau, and the California coast was associated with the ridge along the West Coast, the anticyclonic flow pattern over the West, and the almost complete lack of fronts (fig. 5A).

During the middle ten days of May the circulation pattern had changed in several respects from the first ten days (fig. 3B). One of the most fundamental changes occurred over the Pacific where one major central Pacific trough had replaced the two troughs observed earlier. Accompanying this trough amalgamation were the following changes:

(1) A ridge appeared well off the West Coast, some 15° of longitude west of the previous ridge position along the Coast.
(2) The trough in the Southwest sheared off from its connection to the northeast and built northwestward through Nevada.
(3) A ridge developed over the Southeast and a closed High formed over Florida.
(4) A weak trough developed in the western Atlantic.

Meanwhile at higher latitudes the flow over Canada broke down markedly into a blocking type with closed High centers in western and northeastern sections and a deep Low centered on the southwest side of Hudson Bay. The trough which had been over the Plains extended southward from this Low.

Most of these changes had immediate repercussions on the precipitation regime. As the southwesterly flow off the Gulf of Mexico retrograded so did the general axis of heavy precipitation (fig. 4B). It now stretched from Texas, Oklahoma, and Louisiana northeastward through the Ohio Valley and into northern New England and the Middle Atlantic States. Note how the axis of maximum frontal activity (fig. 5B) shifted to about the same position as the zone of maximum precipitation. Rainfall in excess of 400 percent of normal fell in east Texas and Louisiana in this period leading to the devastating floods mentioned earlier. Meanwhile, rainfall over most of the Southeast decreased to subnormal amounts as the ridge line aloft became more well-defined there and frontal activity decreased markedly. Over the central and northern Plateau and central Rockies precipitation picked up appreciably in connection with the intensifying trough

over Nevada. Heavier-than-normal precipitation generally followed a channel which was located roughly between the zonally oriented 9,900-ft. and 10,000-ft. contours (fig. 3B), south of the axis of maximum number of days with fronts (fig. 5B), and north of the sea level cyclone centers (Chart X). All of the West Coast to the north of central California was wet due to frequent frontal systems moving in from a fast westerly current in the Pacific.

The last ten days of the month saw the changes evident during mid-May culminate in a large amplitude wave pattern over the United States (fig. 3C), which represented a nearly complete change in phase from the pattern which was dominant in the first third. From the central Pacific to eastern North America all waves at middle latitudes had retrograded while the amplitudes of the eastern Pacific ridge, the United States trough and ridge, and the western Atlantic trough all increased. Most striking development over the United States was the establishment in these last ten days of May of a continental anticyclone with center over Alabama. The influence of this huge anticyclonic circulation was remarkable in bringing a complete halt to rainfall over virtually all of the South from Texas and Oklahoma eastward to the western Carolinas (fig. 4C). East Texas, Louisiana, Mississippi, and Tennessee experienced the most extreme change in precipitation regime as rainfall dropped from values in excess of 200 percent of normal during the middle third of the month (fig. 4B) to no rain at all in the last ten days (fig. 4C). This total lack of precipitation obviously brought an end to the severe storminess that had plagued the South in the first two-thirds of the month and it can be seen in figure 1 that no tornadoes were reported in the South after the 19th.

Heavier-than-normal precipitation in the last third of May occurred along the West Coast and the northern Plateau region in association with the deep trough along the Coast. Precipitation in excess of normal also streaked eastward across the northern border region of the United States and then mainly southeastward into the Middle Atlantic States. It is noteworthy that the band of above-normal precipitation across the northern United States was generally oriented in the direction of the flow pattern of figure 3C. Strongest wind speeds at 700 mb. across the United States at this time were located between the 10,000-ft. and 10,400-ft. contours, with the maximum wind axis very close to the 10,200-ft. contour. Note that heavy precipitation fell on the north side of the maximum wind axis from the West Coast to Michigan and generally directly under the maximum winds from Lake Erie to coastal New Jersey and Delaware. It is interesting that the axis of maximum frontal activity in these last ten days (fig. 5C) did not coincide with the axis of heaviest precipitation as it did in the earlier two decades of the month. As an extreme example, fronts were found in eastern Colorado and western Kansas on every one of the last ten days of May, yet precipitation was largely subnormal. The most likely explanation for this lack of precipitation where fronts were most frequent lies in the prevailing upper-level anticyclonic circulation over these fronts which suppressed the upward motion necessary for rainfall, and which also prevented much flow of tropical air over the fronts in mid-troposphere.

TEMPERATURE IN RELATION TO MEAN CIRCULATION

This month's mean temperature anomalies (Chart I–B) were well related to the monthly mean 700-mb. circulation pattern (fig. 2). Temperatures averaged below normal over the western half of the Nation, where the major mean troughs were located and where 700-mb. heights were below normal. The coldest weather (6°–9° F. below normal) occurred over the Great Basin, directly under the negative height anomaly center at 700 mb. The air masses contributing to most of these subnormal temperatures were predominantly polar Pacific, coming into the West under the influence of the prevailing cyclonic circulation which prevented rapid low-level warming of the air mass over the continent. Over the eastern half of the country May was a rather warm and often humid month. This warmth was associated with southerly flow relative to normal over the Mississippi and Ohio Valleys and positive 700-mb. height anomalies in the ridge over the East. Moist maritime tropical and considerably modified polar air were dominant over this part of the country.

Near the boundary between above and below normal monthly mean temperatures there are often marked swings in temperature during short periods within the month. This was especially true this month in the region from the Southern Plains northeastward toward the Great Lakes, where temperatures averaged rather close to normal for the month as a whole (Chart I–B). A striking example of this situation occurred in Oklahoma and northwest Texas where, during the week ending May 18, temperatures averaged 12° to 13° F. below normal when a strong outbreak of cold Canadian air moved southward through the Plains. A week later in the same area temperatures averaged more than 9° to 11° F. above normal as the warm anticyclone developed over the area aloft. Record maximum-temperatures for so early in the season occurred at several stations. Among them was Amarillo, Texas, where 102° F. was reported on the 25th.

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SOME EFFECTS OF THE EVAPORATION OF WIDESPREAD PRECIPITATION ON THE PRODUCTION OF FRONTS AND ON CHANGES IN FRONTAL SLOPES AND MOTIONS

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INTRODUCTION

Surface weather maps in the United States have undergone a succession of dramatic changes in appearance during the past 15 years as one idea after another has dominated the minds of the analysts. The first major change came with the introduction of air mass analysis. After this type of analysis had been stressed for some five years, analysts not only accepted it wholeheartedly but in their enthusiasm, they reached the "spoked wheel" stage of frontal analysis. Out of every cyclone radiated a multitude of fronts, so many as to resemble the spokes of a wheel. Gradually, this first enthusiasm abated, spurious fronts were eliminated, and frontal analysis reached its stage of logical moderation in the United States.

But the stability of frontal analysis was soon upset by a new enthusiasm—this time for "instability lines," "squall-lines," or whatever the analyst chooses to call organized bands of bad weather ahead of the cold front in the warm sector. Since the concept of organized bands of thunderstorms and showers in the warm sector constitutes quite a departure from the ideas of the Norwegian school of air mass analysis, analysts were at first reluctant to accept it. Many forecasters still introduced surface fronts or cold fronts aloft to account for all organized lines of showers. But as time went on, the pendulum swung the other way; all lines of thunderstorms tended to be designated as instability lines, *even those which were fronts*. Unlike the fronts, which are inevitably tied to their air masses, these new instability lines are unrestrained; they move forward or backward, accelerate, decelerate, dissipate, re-form, cause tornadoes or other severe weather. In fact, as explanatory causes of bad weather they are extremely versatile. Meteorologists now appear to be in a stage of enthusiasm for instability lines similar to the "spoked wheel" stage of frontal analysis (fig. 1).

Especially with summer cold fronts it has become the custom with many analysts to draw an instability line at the forward edge of the transition zone and a cold front at the rear edge of this zone. The cold front is, then, coincident with the rear edge of the rain area, so placed because of the dew point discontinuity which necessarily exists between the rainless and the rain areas.

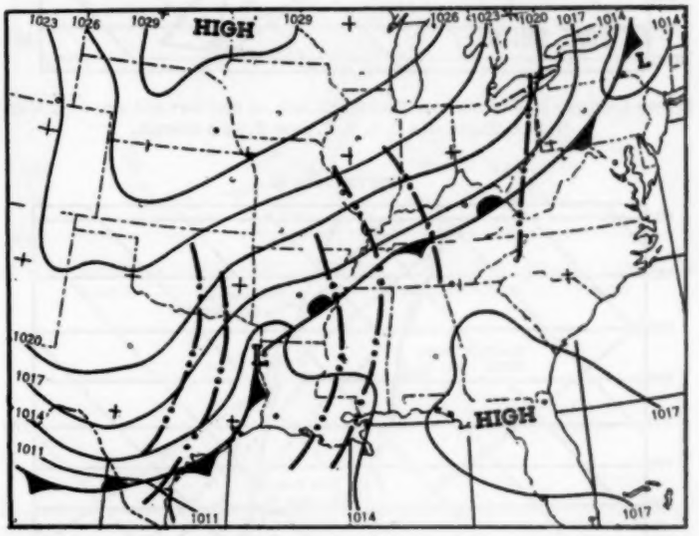

FIGURE 1.—Surface weather chart for 1830 GMT, May 13, 1953, showing what might be termed "over-enthusiam" for instability lines.

This procedure is tantamount to a complete revision of the definition of fronts, since by definition a cold front lies at the *forward* edge of the transition zone, and such revision nullifies many of the forecasting rules based on the classical frontal theory.

What is needed now is a careful study of the relationship between warm season fronts and instability zones, in order to differentiate between the two. The purpose of this paper is to present a case study of a warm season frontal system, one which might easily be confused with an instability line, and to suggest how rainfall can so modify the temperature difference across a front as to cause an acceleration of its movement toward the warm air and to give it many of the characteristics now attributed to instability lines.

The extent to which rainfall can modify the temperature difference across a front increases with the dryness of the air mass into which the rain falls. In fact, the only limit on the amount of cooling which rain can produce is the wet-bulb temperature. In areas where persistent or heavy rain does occur, this limit is reached (see fig. 2). In the case to be described in detail, the difference between the dry-bulb and wet-bulb temperatures was greatest in Texas. It is here that the greatest temperature drop

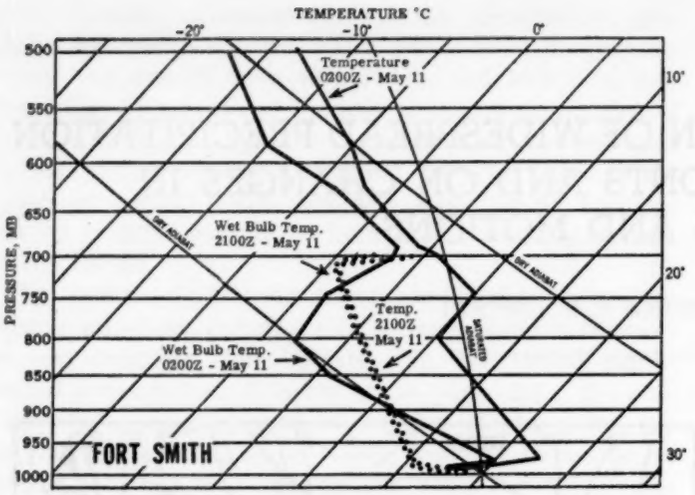

FIGURE 2.—Upper air sounding over Fort Smith, Ark., at 0200 GMT and 2100 GMT, May 11, 1953, plotted on a U. S. A. F. skew T, log p diagram.

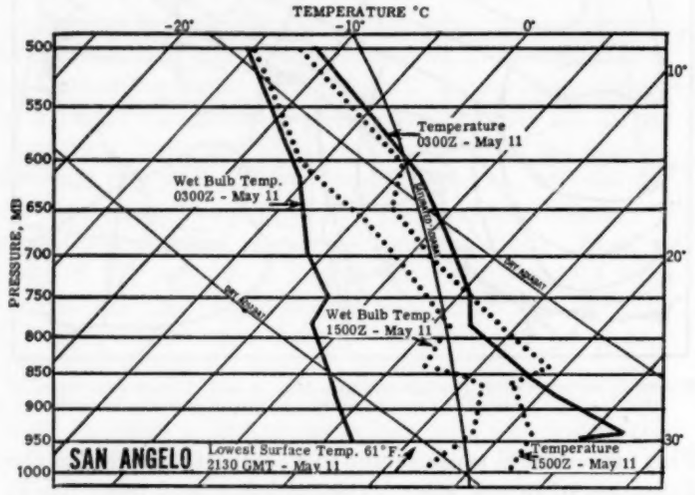

FIGURE 3.—Upper air sounding over San Angelo, Tex., at 0300 GMT and 1500 GMT, May 11, 1953.

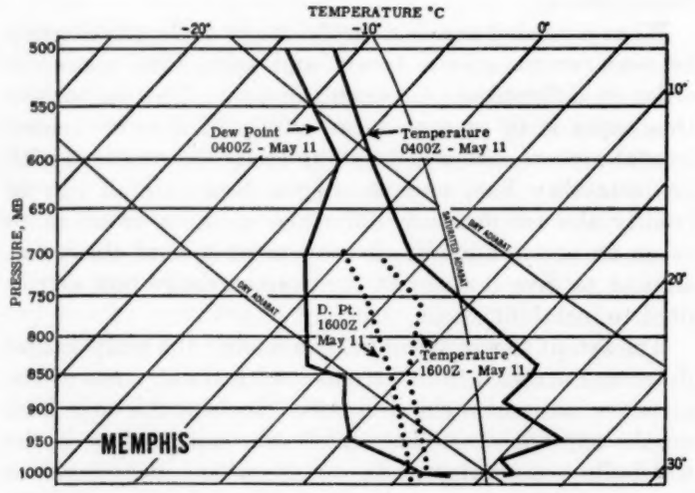

FIGURE 4.—Upper air sounding over Memphis, Tenn., at 0400 GMT and 1600 GMT, May 11, 1953.

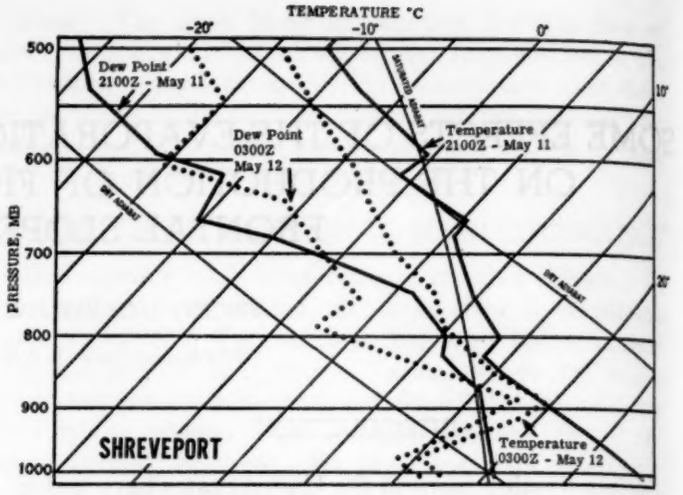

FIGURE 5.—Upper air sounding over Shreveport, La., at 2100 GMT, May 11 and 0300 GMT, May 12, 1953.

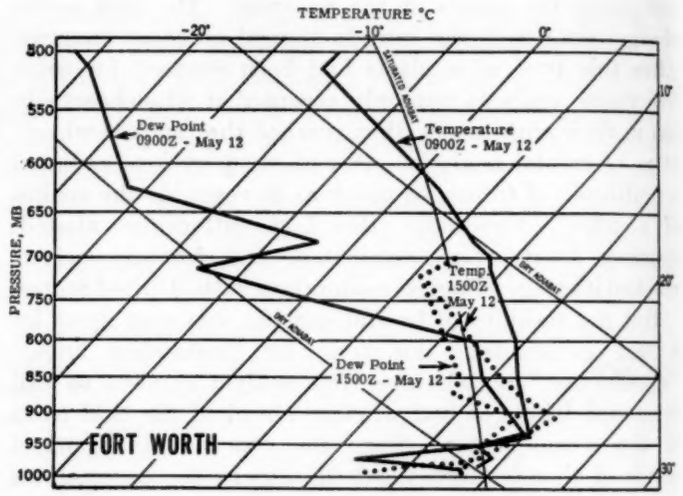

FIGURE 6.—Upper air sounding over Fort Worth, Tex., at 0900 GMT and 1500 GMT, May 12, 1953.

occurred at the surface (fig. 3), accompanied by severe rain and hail storms and several tornadoes (including those which wrecked parts of Waco and San Angelo).

ANTECEDENT GENERAL CIRCULATION

Before presenting the specific case study, we shall describe the sequence of events which generally brings into juxtaposition a cool dry air mass and a warm moist one, requisite for producing the type of frontogenesis and frontal accelerations that is dealt with here.

To start with, a rather deep cold air mass enters the West Coast of the United States from the Pacific. As it crosses the western third of the country, the southern portion of this air mass is rather rapidly heated over the southwestern desert region and is further warmed by downslope motion east of the Continental Divide so that when it approaches the tropical maritime air moving northward from the Gulf of Mexico, the temperature con-

trast between these two air masses has become negligible. This means that the frontal slope will be very steep, approaching the vertical as the two air masses approach the same density. At this stage, there is little evidence of the front in the temperature field on the sea level, 850- and 700-mb. charts, although it is usually possible to find a moderate temperature gradient between the colder mP air and mT air at 500 mb. At 850 and 700 mb. the most obvious difference between these two air masses is the moisture content, which is usually very marked indeed. This moisture difference does account for a difference in the virtual temperature (and density) so that the slope to the front is slightly less than vertical. The front is generally oriented now from northeast to southwest and exhibits little motion, although it may start to move northward as a weak warm front.

HYPOTHESIS

Suppose that at this point, when the frontal motion is slight and the frontal slope is nearly vertical in the lower levels, the warm air overruns the front and heavy thunderstorms break out north of the front. Because of the excessive dryness of the "cold" air mass into which the precipitation falls, evaporation and consequent cooling of this air mass will be very rapid. Within an hour or two, the temperature difference across the front may increase greatly throughout a layer as deep as 10,000 feet at times.

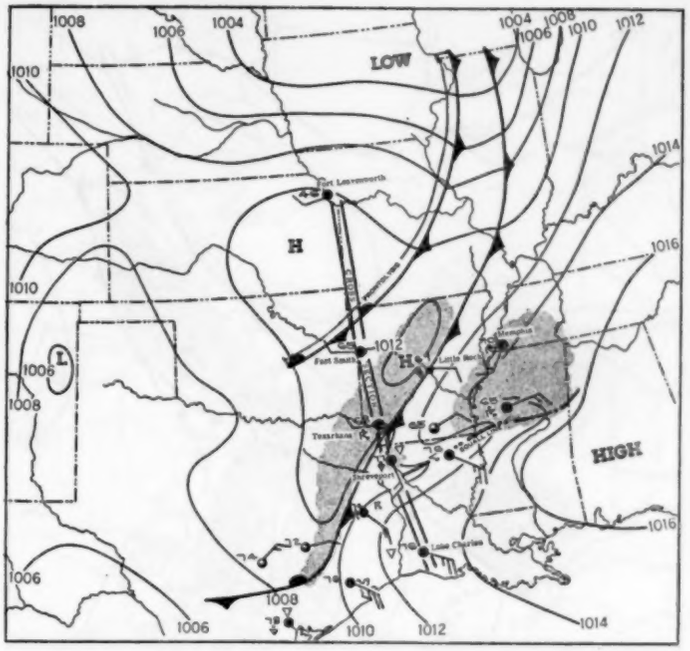

FIGURE 7.—Surface weather chart for 0330 GMT, May 11, 1953. Shading indicates the areas of active precipitation. Figure 8 is a corresponding cross-section.

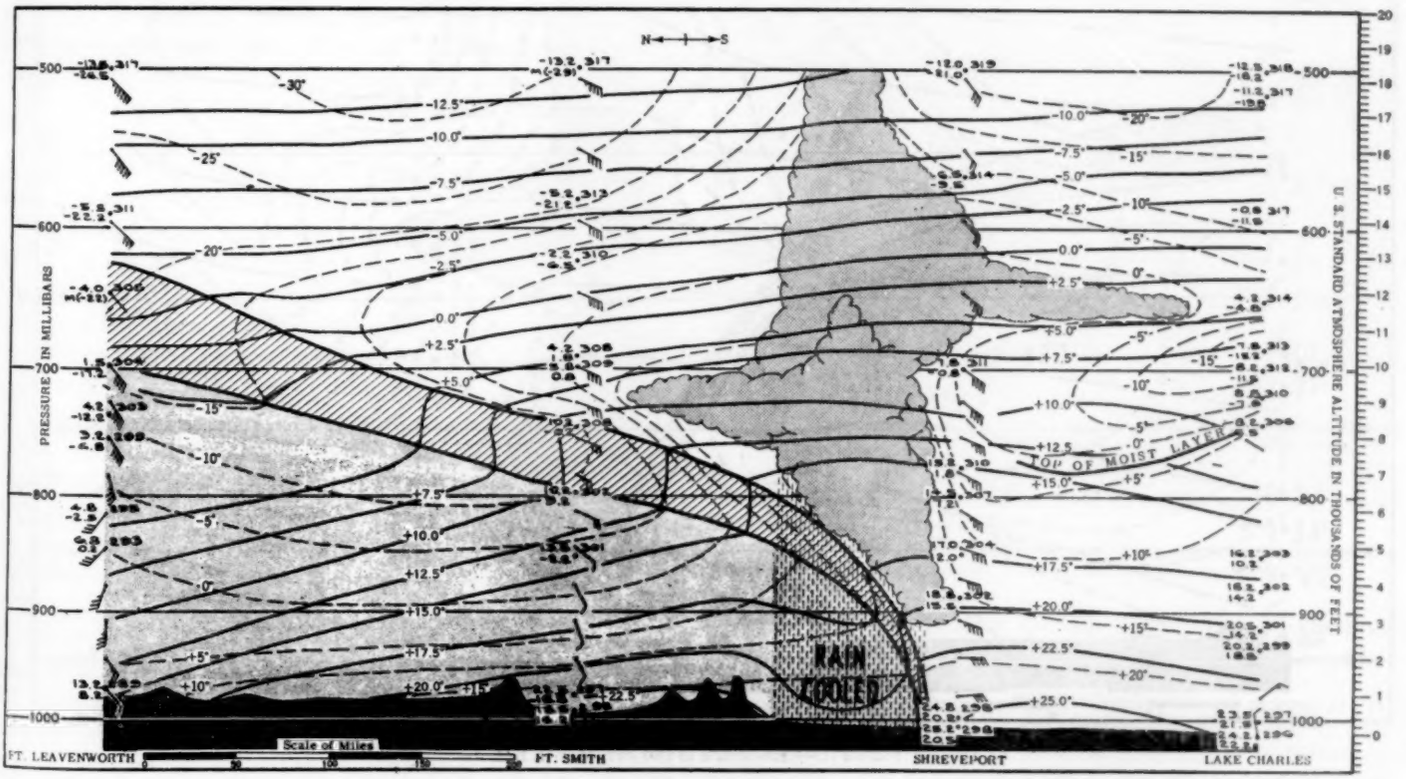

FIGURE 8.—Cross-section at 0300 GMT, May 11. Isotherms in °C. are the solid lines and isodrosotherms are shown as dashed lines. The plotted data also include potential temperatures (not analyzed). The heavy solid line indicates the transition zone of the front. Winds are plotted with reference to the directions (along the top of the diagram) which show also the orientation of the stations.

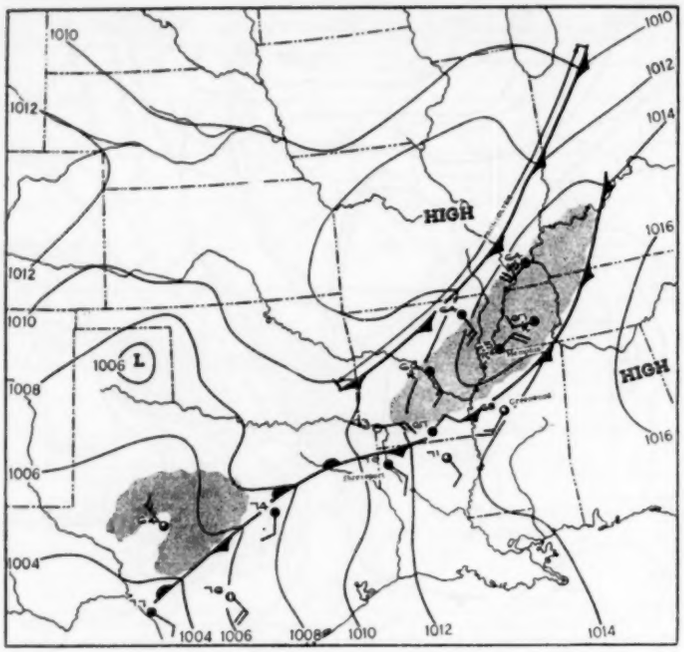

FIGURE 9.—Surface weather chart for 1230 GMT, May 11, 1953.

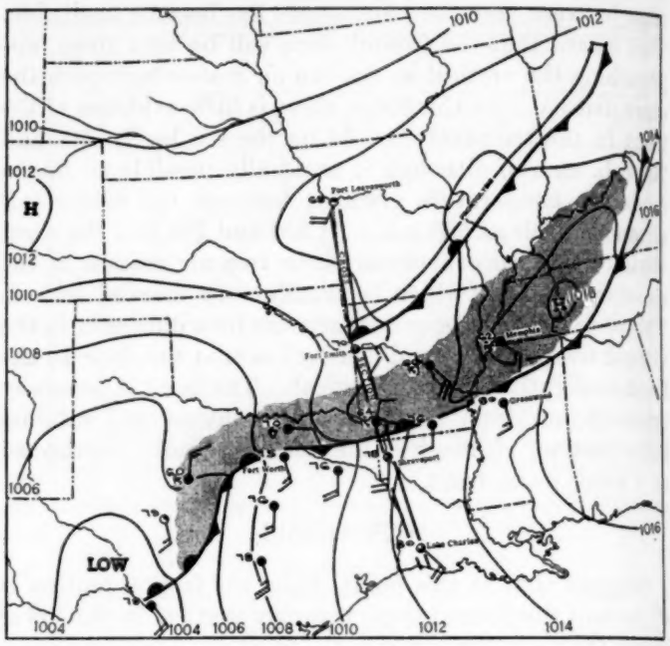

FIGURE 10.—Surface weather chart for 1530 GMT, May 11, 1953. Figure 11 is a corresponding cross-section.

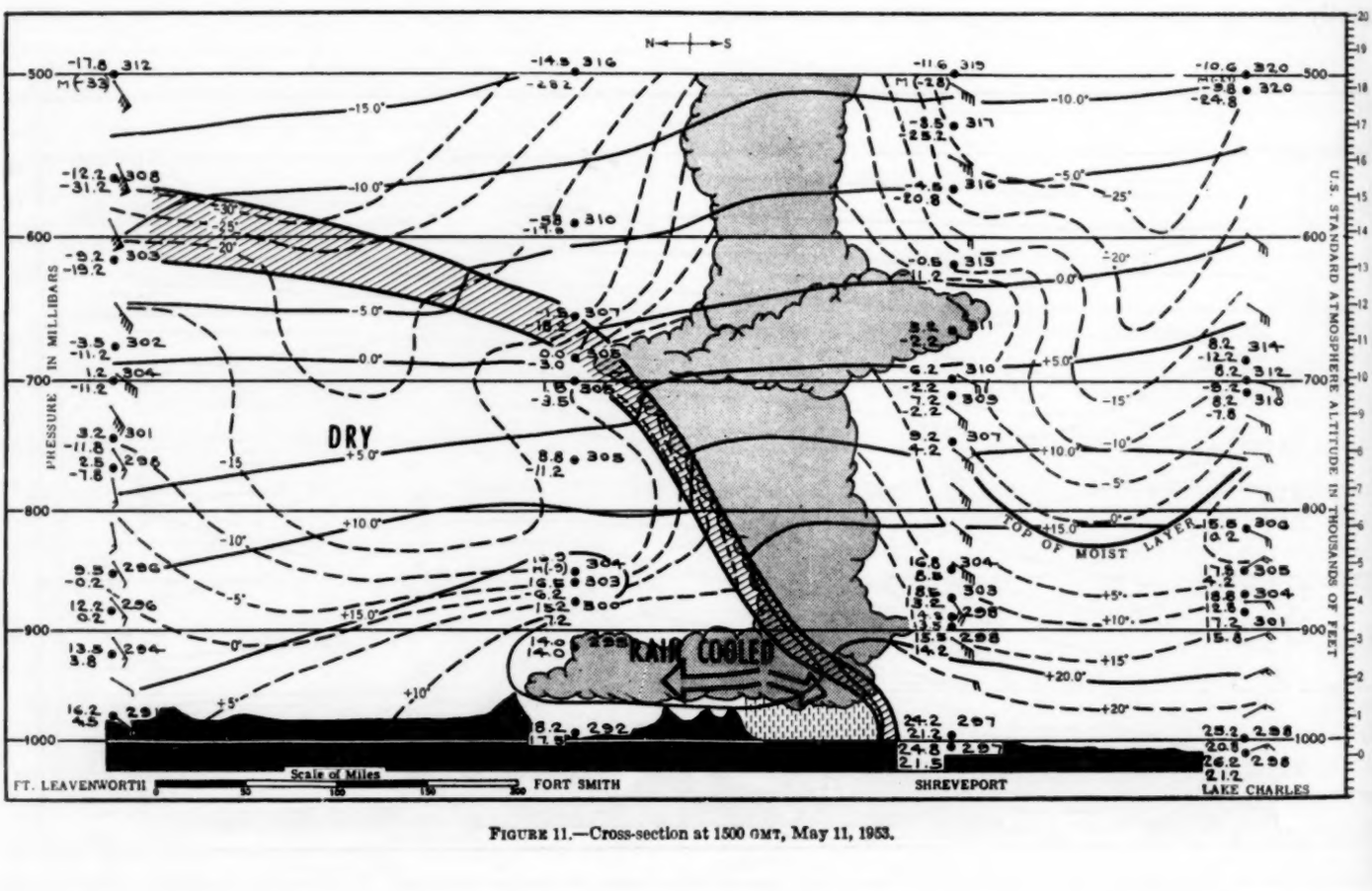

FIGURE 11.—Cross-section at 1500 GMT, May 11, 1953.

This obviously changes the pressure distribution [1, 2]; there is now a ridge forming north of the front in the rain-cooled air. With this rain-induced High and with the pronounced temperature gradient, it is hypothesized that the steep slope of the front can no longer be maintained and the cold air mass must flatten out with the front readjusting its slope to the new temperature difference between the air masses (see the following article by Wexler [3]). If it is assumed that at the top of the evaporation-cooled layer there is no appreciable change in temperature, the readjustment of the frontal slope requires that the surface front suddenly move from the rain-induced high-pressure area as the cold air mass spreads out. The effect of evaporational cooling on the frontal motion is most dramatic when a front which has been moving northward as a weak warm front suddenly reverses its motion and surges southward as a strong cold front.

Such a surge of the cold front may prove to be the trigger which frequently sets off pressure jumps. It has been suggested that pressure jumps can be produced by the sudden acceleration of part of a cold front [4, 5]. It seems that just such frontal accelerations may be easily produced by the process of evaporational cooling outlined above.

THE CASE OF MAY 9–13, 1953

We turn now to the detailed example, which deals with a storm affecting the midwestern part of the United States from May 9–13, 1953, producing numerous tornadoes, widespread heavy rains, and rain-cooled air masses, some of which intensified existing fronts and some of which produced frontogenesis. A series of hourly maps (of which every third map is reproduced) was prepared to study the details of the movements of the fronts and the precipitation areas. Also prepared was a group of soundings at selected stations (figs. 2–6), showing the amount of cooling produced by the rain.

The first surface chart (fig. 7), synoptic with the cross-section of figure 8 and with the sounding of figure 2, shows a front, quasi-stationary in the south, oriented from northeast to southwest, and extending from Illinois to Texas. This front separates dry modified Pacific air to the northwest from an mT air mass to the southeast. An important feature of this situation was the structure of the mT air mass. It was much more unstable than usual, i. e., above the lower moist layer, there was aloft unusually dry air with a steep lapse rate. (The Showalter stability index [6] was −4.) This meant that the instability of the mT air could be released by a small amount of lifting. Some of this instability had already been released in the warm sector along the instability line through northern Louisiana and central Alabama (fig. 7). Frontogenesis was occurring along the instability line at this time due to convergence between the air flowing outward from the rain-cooled portion of the warm sector and the unmodified warm air to the south. Had the rain continued north of the zone of frontogenesis, an important front might have formed. The rain, however, died out during the night, ending the frontogenesis.

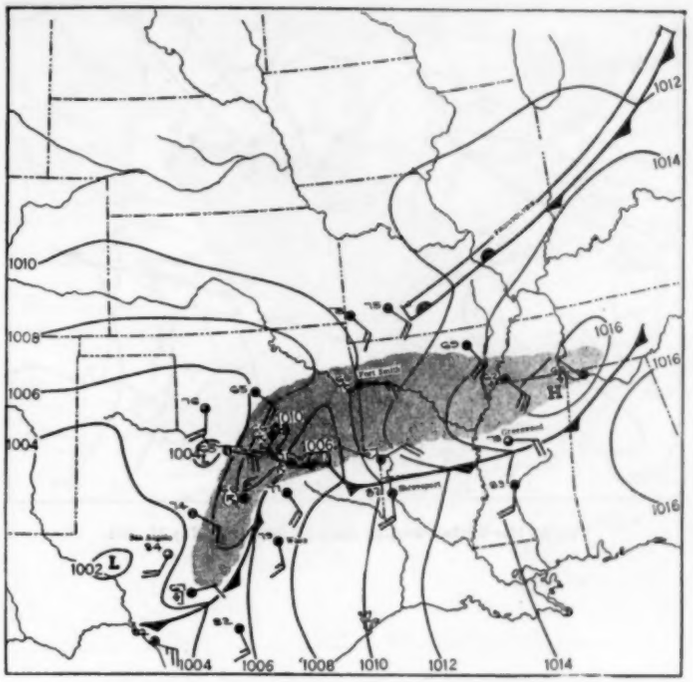

FIGURE 12.—Surface weather chart for 1830 GMT, May 11, 1953.

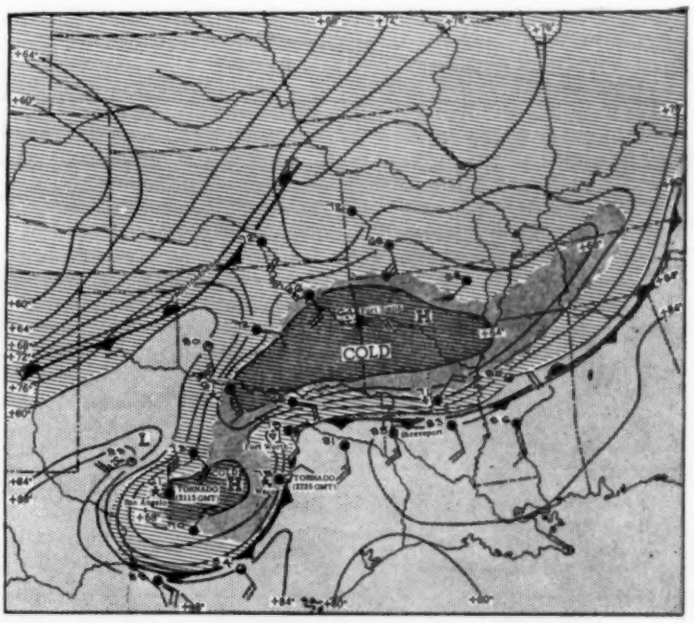

FIGURE 13.—Surface weather chart for 2130 GMT, May 11. Surface isotherms are for every 4° F. Note the occurrence of tornadoes at San Angelo and Waco, Tex.

Behind the major cold front, on the other hand, the rain persisted with rapid cooling due to evaporation in the dry air. By 1230 GMT, May 11 (fig. 9) the rain-induced High was already distinctly formed in western Tennessee

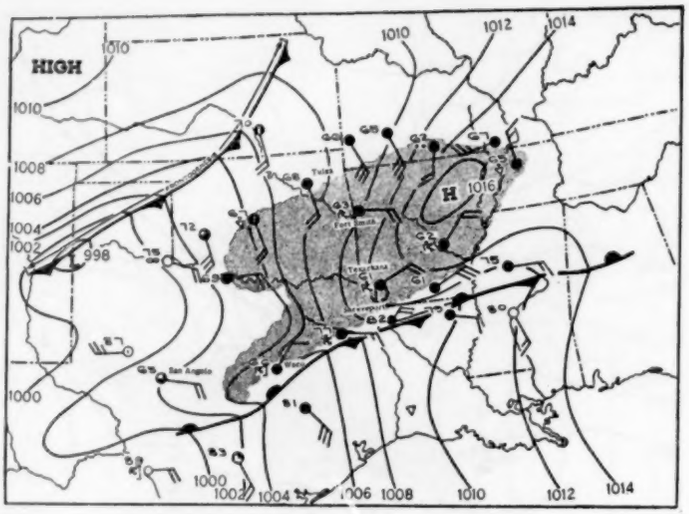

FIGURE 14.—Surface weather chart for 0030 GMT, May 12, 1953.

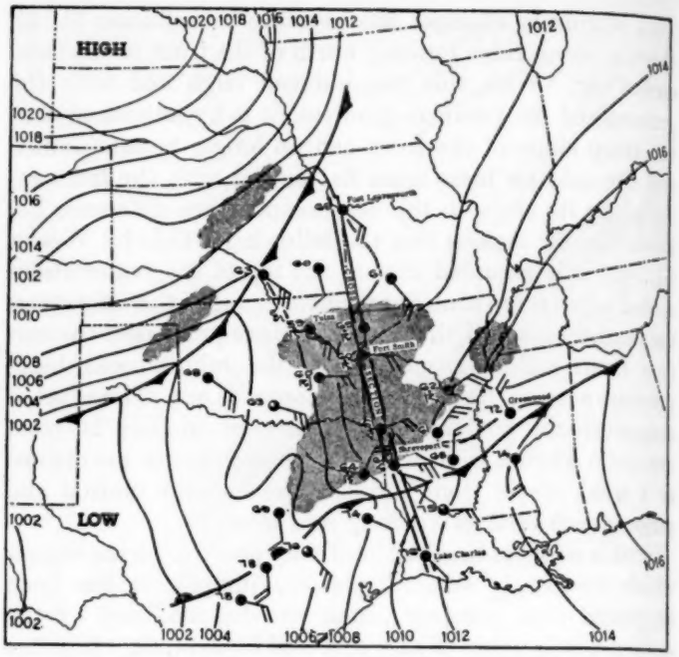

FIGURE 15.—Surface weather chart for 0330 GMT, May 12, 1953. Figure 16 is a corresponding cross-section.

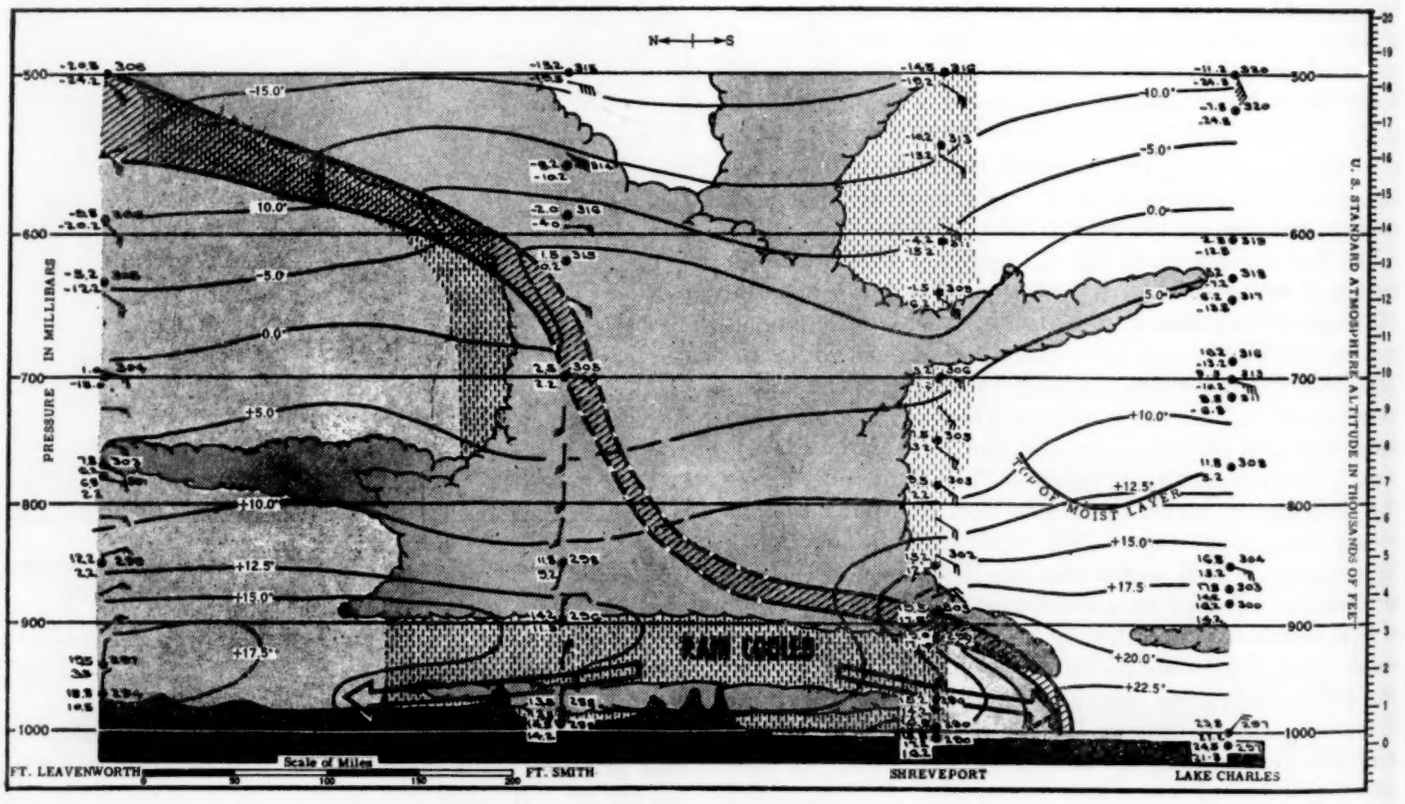

FIGURE 16.—Cross-section at 0300 GMT, May 12, 1953.

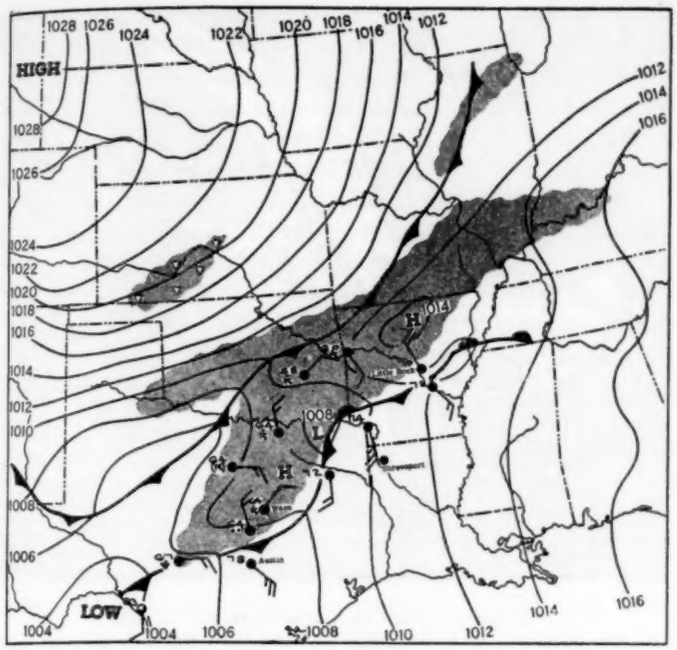

FIGURE 17.—Surface weather chart for 1530 GMT, May 12, 1953.

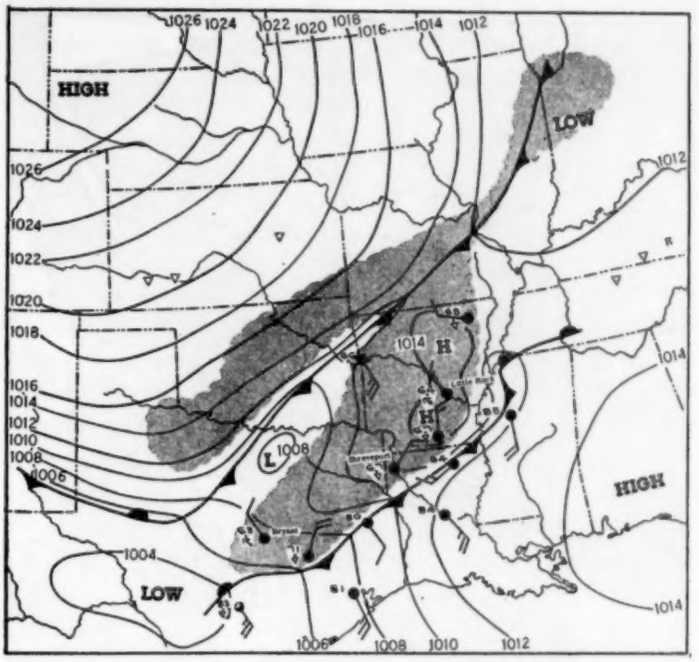

FIGURE 19.—Surface weather chart for 2130 GMT, May 12, 1953.

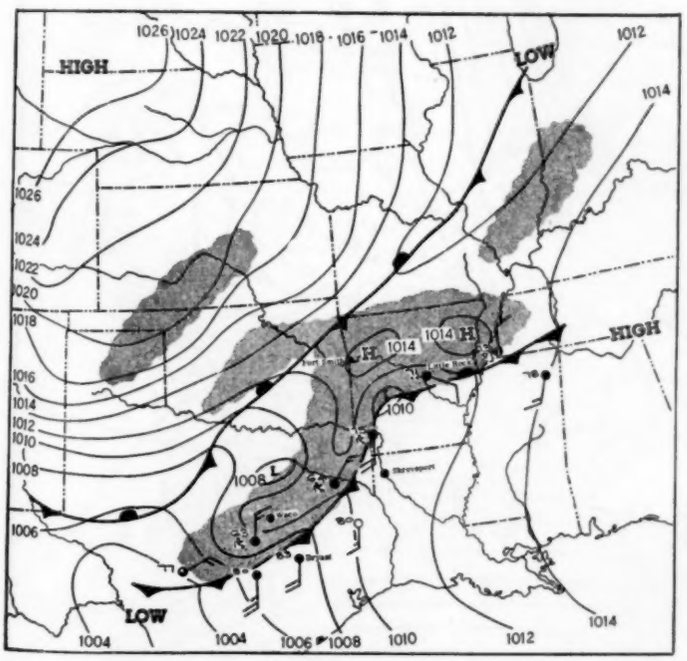

FIGURE 18.—Surface weather chart for 1830 GMT, May 12, 1953.

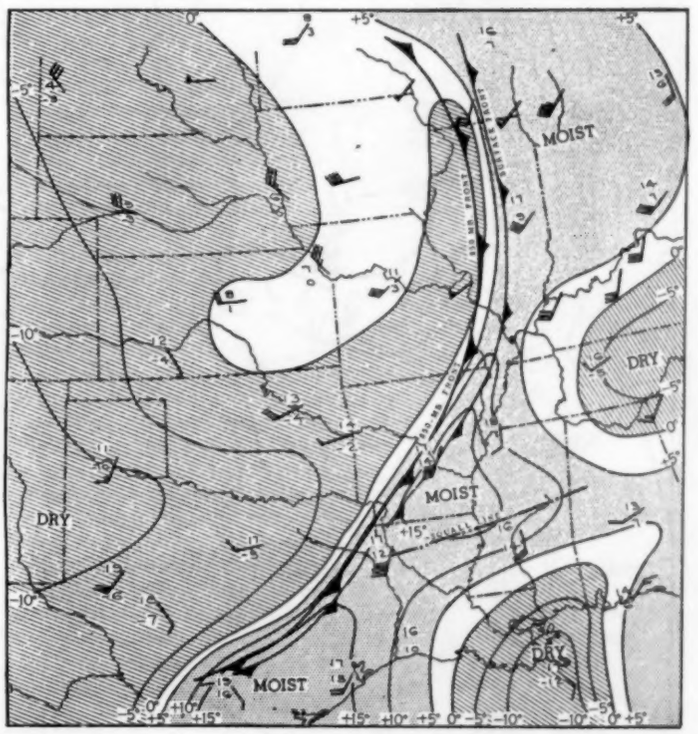

FIGURE 20.—850-mb. chart for 0300 GMT, May 11, 1953, showing moisture distribution. Isodrosotherms are for every 5° C. The data show the winds and temperatures as well as dew points.

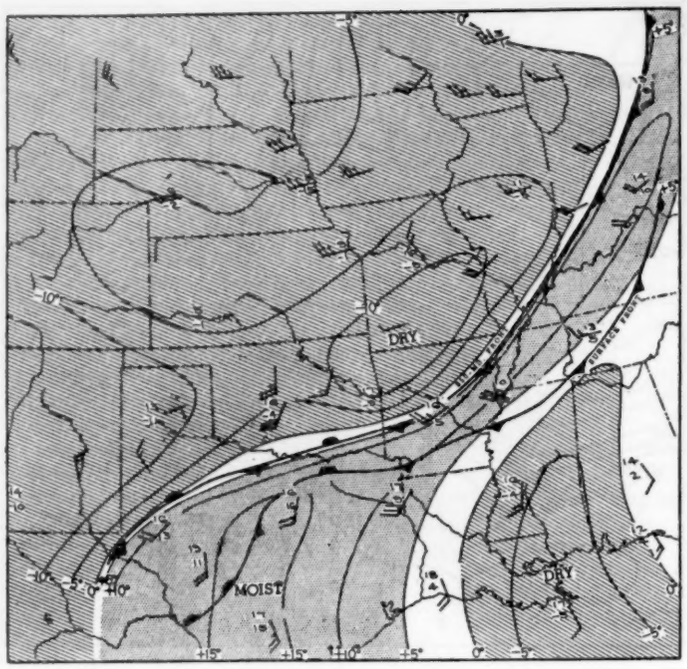

FIGURE 21.—850-mb. chart for 1500 GMT, May 11, 1953.

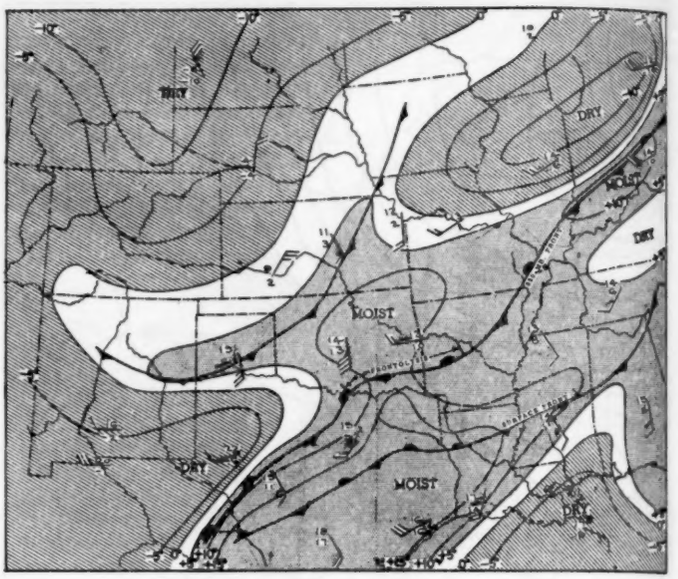

FIGURE 22.—850-mb. chart for 0300 GMT, May 12, 1953.

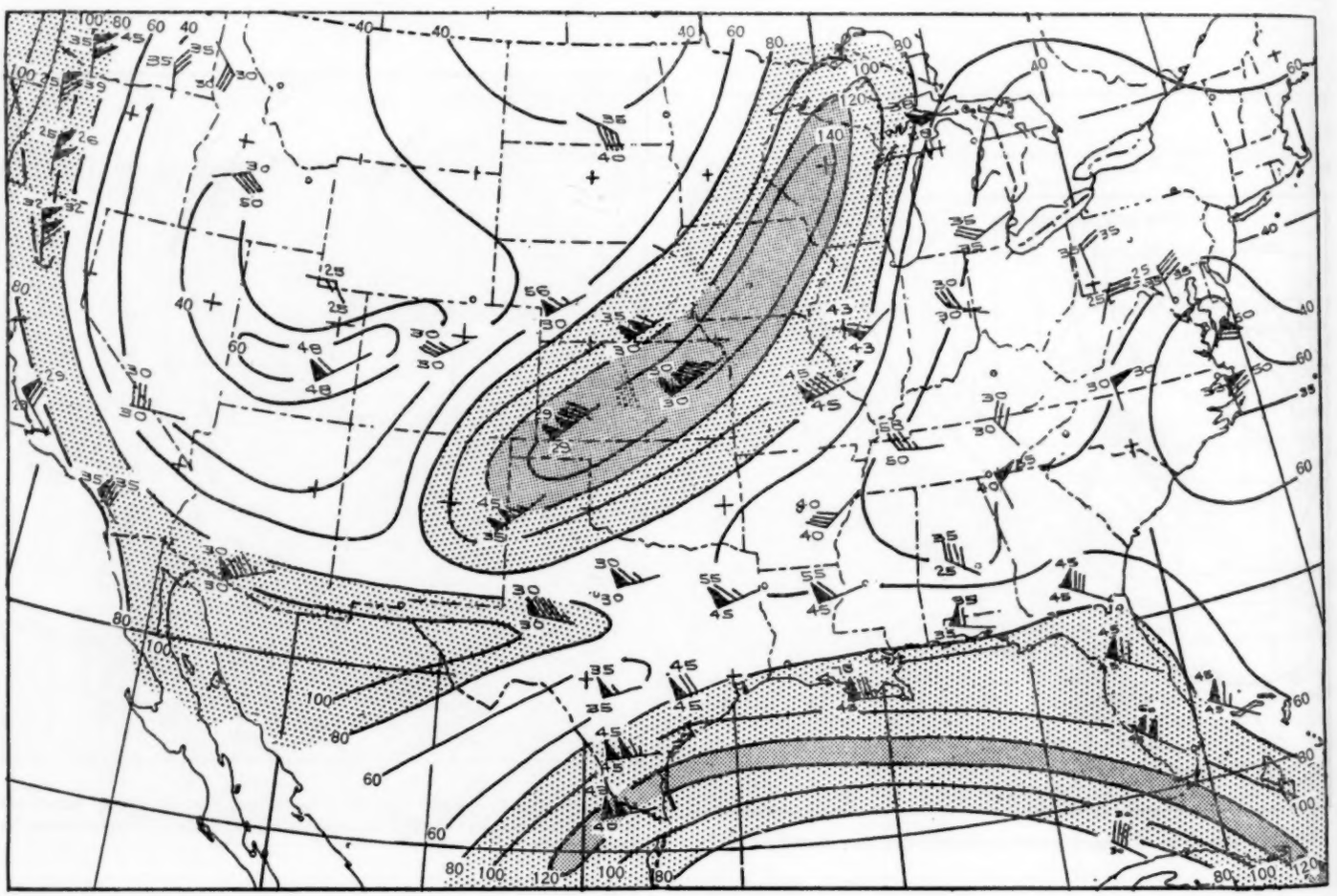

FIGURE 23.—Maximum-isotach chart for 0300 GMT, May 11, 1953. A full barb means 10 knots and a flag means 50 knots. The small number at the tip of the shaft indicates the level of the maximum wind in thousands of feet. The number at the tips of the flags and barbs is the highest level which the sounding reached. Isotachs are for every 20 knots.

and the surface winds were blowing out from it in all directions. During the next few hours the thunderstorms continued in this general area, merging with those in Texas (fig. 10). The evaporational cooling increased the temperature gradient across the front (fig. 10) and resulted in a flattening of the frontal slope, accomplished in this case by a southward motion of the surface front (fig. 11). On the ground the front was driven southward by a narrow band of northeasterly winds (see figs. 9 and 10), passing Greenwood, Miss., at 1730 GMT and continuing slowly to the south, despite the fact that during this entire period the general wind flow from about 3000 feet to the stratosphere was opposed to this motion (i. e., from the southwest).

A second southward surge of the rain-cooled air mass occurred in southwestern Arkansas (figs. 12–15). The soundings for Fort Smith, Ark. (fig. 2), best illustrate the cooling due to the rain which was instrumental in producing the change in frontal slope and this second southward surge of the front. The first sounding (0200 GMT, May 11), taken before the rain began at Fort Smith, shows the extreme dryness of the lower portion of the cold air mass north of the front. At this time the overrunning moist air was just reaching Fort Smith at about 10,000 feet. The sounding 19 hours later was taken when a thunderstorm was in progress at the station. Again, referring to the sea level charts near Fort Smith for this period, as the rain area increased in size and intensity in southwestern Arkansas, the sea level pressures rose as widespread cooling occurred (note the sea level isotherms in fig. 13 and the isotherms in low levels in fig. 16), surface winds became northeasterly, and the front which had been nearly stationary between Texarkana, and Shreveport, accelerated rapidly to the south (see figs. 11–13). It was during this period that devastating tornadoes struck Waco and San Angelo, Tex.

A third major surge of the cold front took place the following afternoon, again in southern Arkansas and northern Louisiana. By May 12, 1530 GMT (fig. 17) widespread severe thunderstorms again occurred north of the front. Again the rain, falling through the initially dry air mass, evaporated; the air mass was cooled, the frontal slope decreased, and the activated cold front surged to the south. As the cold air spread out, it lifted the overrunning warm air and set off additional thunderstorms. This is illustrated by the 1530—2130 GMT charts for May 12, 1953 (figs. 17–19).

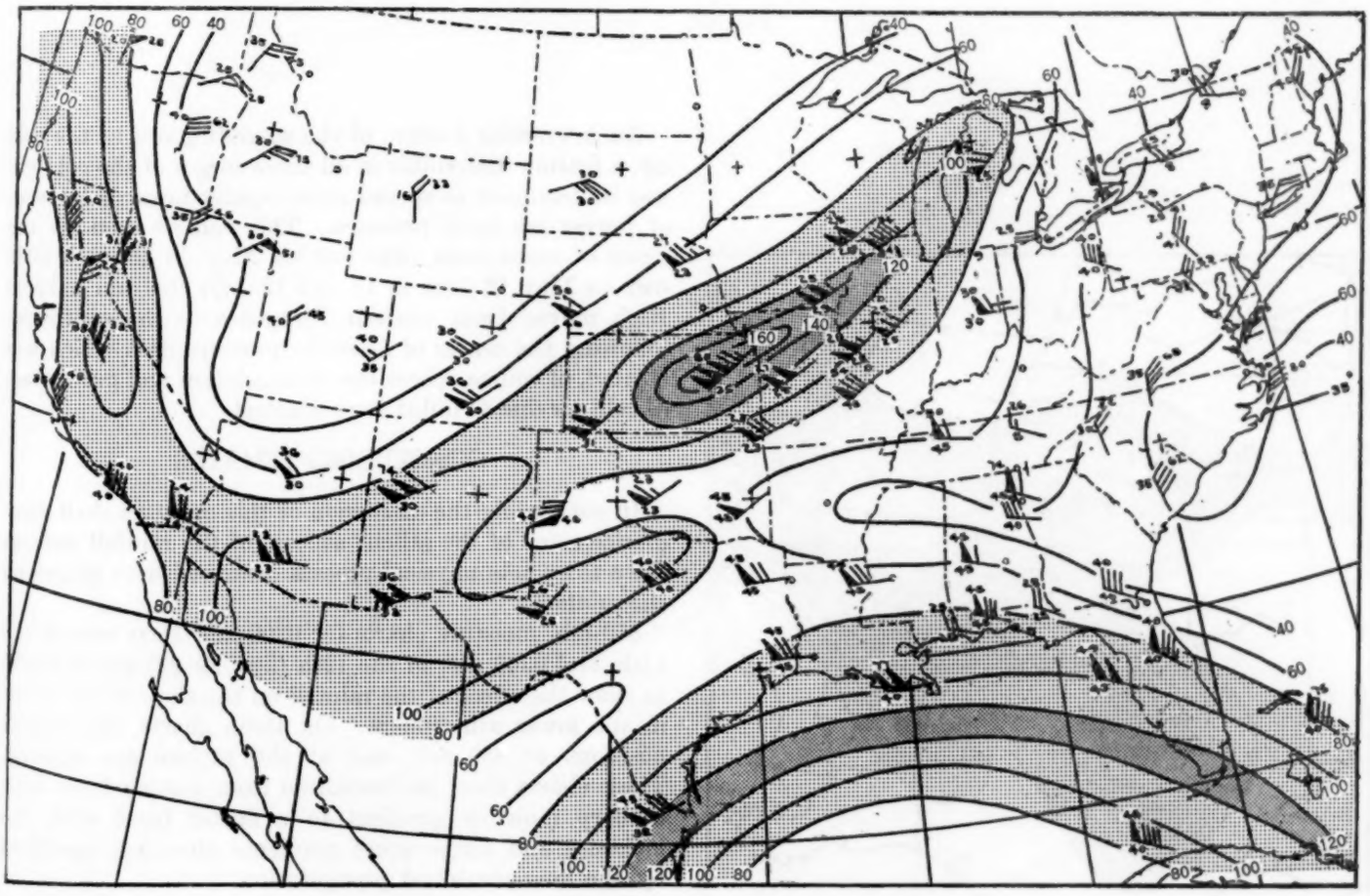

FIGURE 24.—Maximum-isotach chart for 1500 GMT, May 11, 1953.

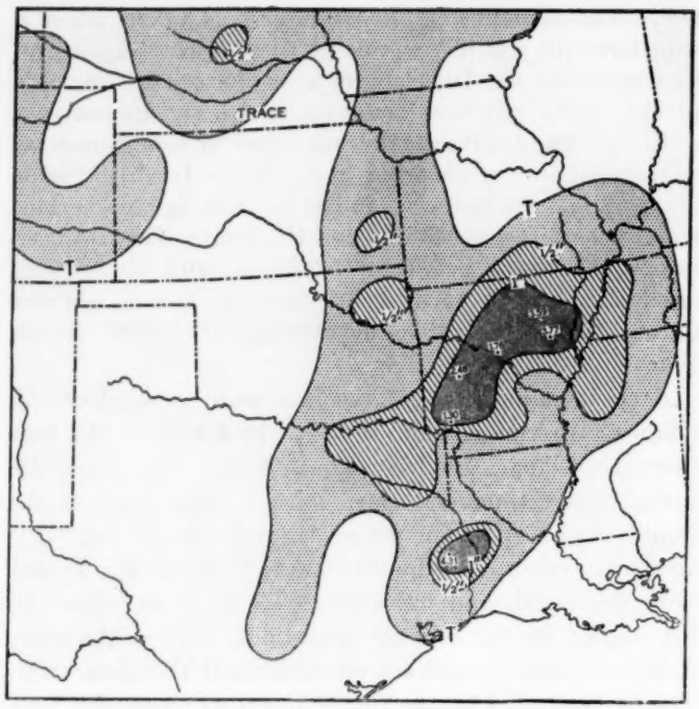

FIGURE 25.—24-hour precipitation chart ending at 1230 GMT, May 11, 1953. Some of the amounts in excess of one inch are shown.

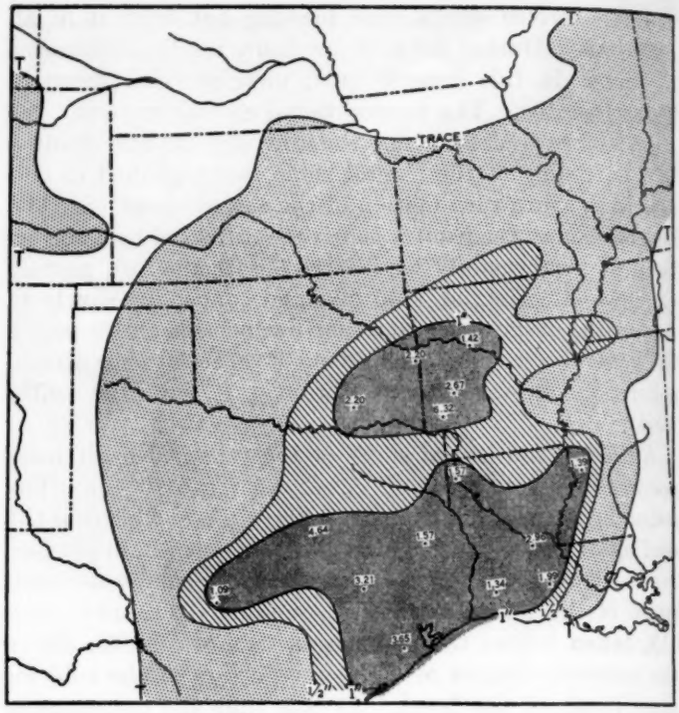

FIGURE 27.—24-hour precipitation chart ending at 1230 GMT, May 13, 1953.

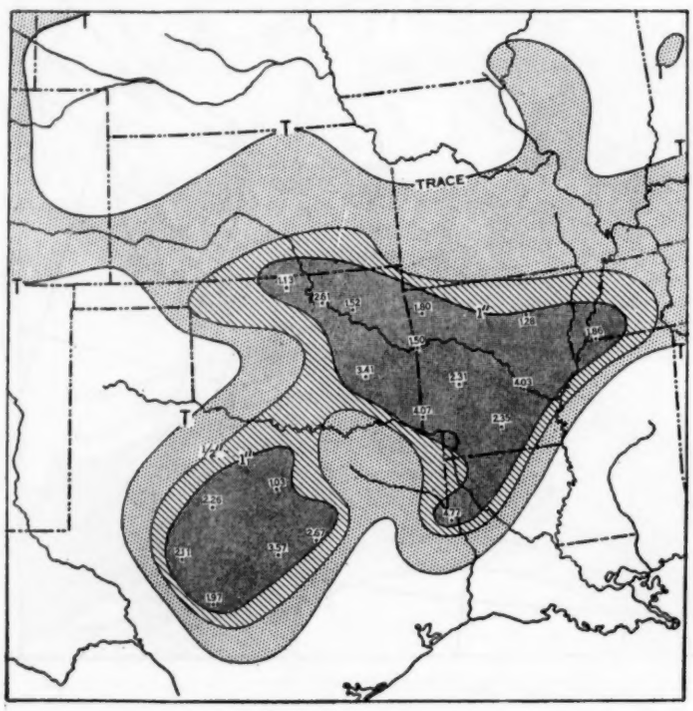

FIGURE 26.—24-hour precipitation chart ending at 1230 GMT, May 12, 1953.

An interesting feature of the spreading out of the cold air, a feature discernible in all three surges of the cold air, was its tendency to spread most rapidly towards the area of lowest sea level pressure. This can be seen on the series of maps from 1230 GMT on May 11 through 0030 GMT on May 12 (figs. 9, 10 and 12–14): the rain induced High moved from western Tennessee to the southwest following the center of 3-hourly pressure rises which was located, of course, where the accumulation and production of cold air due to influx were greatest.

CONCLUDING REMARKS

In concluding the discussion of this case, we shall summarize some of the salient aspects of the rainfall pattern since the whole physical process described here hinges on the rainfall.

850-mb. moisture charts for 0300 and 1500 GMT of the 11th, and 0300 GMT of the 12th (figs. 20–22) are included to show the effect of the rainfall on the slope of the front in the lower atmosphere. On these charts the frontal positions at 850 mb. and at the surface are entered. These charts show the transition from a steep front with a sharp moisture gradient to a flatter front with the moisture and temperature gradients obviously modified by the long-continued precipitation.

Two maximum-isotach charts, showing the location of the branches of the jet stream for 0300 and 1500 GMT, May 11 (figs. 23 and 24), are included here because of the

intimate causal relationship which often exists between the jet and a narrow band of precipitation [7]. It is evident from the charts presented here that neither branch of the jet lies directly over the rainfall area in the south central United States. However, the chart for 1500 GMT shows that one maximum isotach was approaching the rain area at that time. In this case the relationship between the jet and the narrow band of precipitation is inconclusive, despite the suggestive elongated shape of the rainfall area.

Finally, 24-hour rainfall maps for May 11, 12, and 13 (ending at 1230 GMT) are included (figs. 25–27). Besides providing a graphical summary of the precipitation pattern for the period we have been discussing, these charts show that the area of maximum rainfall is coincident with the areas on the ground where maximum cooling and spreading out of the cold air took place. This close relationship between the rainfall pattern and the motion of the cold air lends credence to the original hypothesis: that warm-season precipitation falling into dry air behind a front can cause sudden accelerations of the front and endow it with attributes now generally associated only with instability lines.

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3. H. Wexler, "Readjustment of a Front after Cooling by Precipitation," *Monthly Weather Review*, vol. 81, no. 5, May 1953, pp. 152–154.
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READJUSTMENT OF A FRONT AFTER COOLING BY PRECIPITATION

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INTRODUCTION

The problem of the readjustment of a front in unstable equilibrium, such as might be caused by sudden cooling of the cool dry air by precipitation, has been treated by Starr [1]. Two air masses differing in density by 4 percent (~10° C.), initially motionless, and separated by a vertical front, are allowed to come to equilibrium. For the first model, when the two air masses, 8 km. deep, comprise the entire atmosphere, the displacement southward of the front from its original position is 500 km. In the second model, the readjustment of the same two air masses takes place under a deep upper layer of 20 percent lesser density than the lower cool air, and in this more realistic case the displacement southward of the front is 450 km.

The numerical computations are extremely tedious because of the large number of significant figures which must be carried to obtain meaningful results. Hence, it does not appear worthwhile to repeat Starr's calculations for a model where perhaps only the lower 2 km. of the cool air, whose density is initially very nearly equal to that of the warm air, is suddenly cooled by precipitation 5° to 10° C. A simplified model is adopted which lends itself more readily to calculation and gives results comparable to Starr's more complicated and arduous attack.

THE FRONTAL MODEL

In figure 1 we assume that, because of cooling of the cool, dry air by precipitation, the front in the north-south vertical cross-section, originally denoted by the line ABC, changes into the line DC. The problem is to find the displacement southward of the front, δ=AD, and also the kinetic energy liberated by the resulting decrease in potential energy.

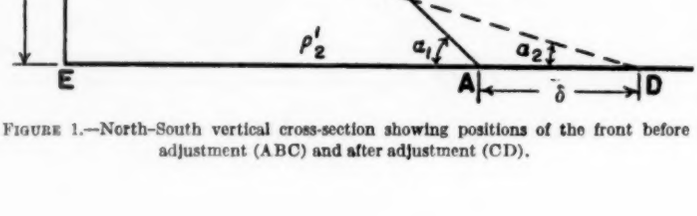

FIGURE 1.—North-South vertical cross-section showing positions of the front before adjustment (ABC) and after adjustment (CD).

The uniform density of the cool, dry air before cooling by rain is ρ_2, after cooling, ρ'_2; the density of the warm air is ρ_1, the original frontal angle, α_1, and the final angle, α_2. H is the undisturbed depth of the cool air mass.

COMPUTED AND OBSERVED DISPLACEMENTS OF FRONT

From the Margules formula for the slope of the front, where it is impliicitly assumed that the frontal slope is much larger than the slope of an isobaric surface, we have

$$(1)\qquad \begin{cases} \cot \alpha_1 \cong -\dfrac{g}{f}\dfrac{\rho_2-\rho_1}{\rho_2\,\Delta u_1}=\dfrac{g}{fT}\dfrac{\Delta T_1}{\Delta u_1} \\ \cot \alpha_2 \cong -\dfrac{g}{f}\dfrac{\rho'_2-\rho_1}{\rho'_2\,\Delta u_2}=\dfrac{g}{fT}\dfrac{\Delta T_2}{\Delta u_2} \end{cases}$$

where

g is the acceleration of gravity

f is the Coriolis parameter

T is a mean virtual temperature of the air masses in °K.

ΔT_1, ΔT_2 are the initial and final virtual temperature differences across the front, and

Δu_1, Δu_2 are the initial and final difference across the front of the wind components parallel to the front.

Conservation of mass of the warm air before and after adjustment leads to

$$(2)\qquad \delta=\frac{H}{2}(\cot \alpha_2-\cot \alpha_1)$$

which with aid of equation (1) reduces to

$$(3)\qquad \delta=\frac{gH}{2fT}\left(\frac{\Delta T_2}{\Delta u_2}-\frac{\Delta T_1}{\Delta u_1}\right)$$

The analysis here, as in Starr's computations, goes from an initial (unstable) state to a final (stable) configuration. A study of the details of the readjustment process is not made although it is clear that an inertial oscillation of the front about its final equilibrium position will result as the front first overshoots, then is forced back, and so on, with decreasing amplitude. The period of an inertial oscillation is one-half a pendulum day, or 24 hours at latitude 30°, and it is likely that, friction neglected, the period of oscillation of the front has approximately the same value. A portion of the potential energy liberated when the frontal slope decreases as a result of precipitation cooling,

must appear both as north-south oscillations of the front and as increased winds; the remainder being dissipated as a result of surface friction.

Referring now to the material in the preceding paper [2], let us see if frontal oscillations occurred and how the computed and observed values of frontal displacement agreed.

In figure 2 are presented two arrow plots showing displacements of the front along meridians 94°W. (roughly Texarkana to Shreveport, La.) and 100°W. (through Abilene, Tex.), as taken from the 3-hourly weather charts, including those shown in [2]. There are not enough stations to guarantee accuracy of these displacements, but at least they represent the best judgment of an experienced analyst taking into account all the available observations.

Both plots in figure 2 show oscillations of the front along their respective meridians. Along the 94th meridian at 0300 GMT on May 11, 1953, the front moves in an apparent damped oscillation until 2100 GMT, with a period of 6 to 9 hours and a maximum displacement of 70 km. southward. Then the front starts another series of oscillations of larger period and amplitude—18 hours and maximum displacement of 330 km. toward the north. This oscillation shows the effect of a cold front passage from the west following the movement eastward of a small wave cyclone along the front. However, before this occurred the total displacement southward from 0630 GMT on the 11th to 0630 GMT on the 12th was 260 km.

To compute frontal displacements from equation (3) we must have values of ΔT_1, ΔT_2, Δu_1, Δu_2. It is difficult to obtain representative values of these virtual temperature and wind differences across the front, especially when the frontal zone is so narrow compared to the distance between aerological stations. As nearly as can be judged from the data, there was initially no difference in *dry-bulb temperature* across the southern portion of the front—see, for example, figure 20 of [2], where at 0300 GMT, May 11, at 850 mb., dry-bulb and dew-point temperatures at Ft. Worth are 17° C., −5° C. as compared to 17° C., 12° C. at Shreveport. On the other hand, the difference in virtual temperature between the two stations is 1.3° C. This is probably larger than the difference right at the front. Accordingly, values of δ will be computed using an initial virtual temperature difference of $\Delta T_1=1°$ C.

To obtain an estimate of the final virtual temperature difference, refer to figures 8 and 16 of [2] where at 1,000 mb. (just above the surface) the dry-bulb temperature at Shreveport, La., dropped 8° C. from May 11, 0300 GMT to May 12, 0300 GMT. Most of this drop is attributed to precipitation cooling of the cool dry air. This value fits in well with an observed 6° C. average cooling in the lowest 3 km. from 0200 to 2100 GMT on May 11 at Fort Smith, Ark., and the 6° to 8° C. wet-bulb temperature depression in the same layer at 0200 on May 11 (see figure 2 of [2]). We shall therefore take 7° as the decrease in dry-bulb temperature, or 6° as the decrease in virtual temperature, induced by evaporation of precipitation falling through the cool air, thus giving $\Delta T_2=7°$ C.

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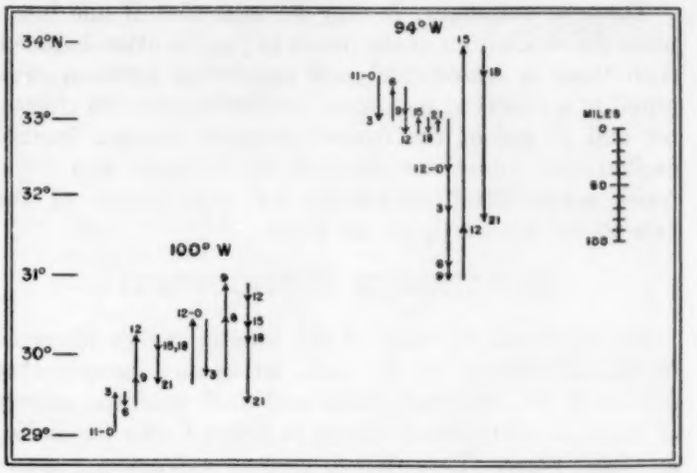

FIGURE 2.—Frontal oscillations along meridians 94° W. and 100° W. during period 0030 GMT May 11 to 2100 GMT May 12. Mileage scale applies only to distance along the meridians. Numerals at arrow-ends refer to dates and times (GMT).

Estimates of representative wind differences across the front are even more difficult to make. At the 850 mb. surface for 0300 GMT, May 11 (figure 20 of [2]) the winds at Ft. Worth and Shreveport are from the southwest, 10 knots and 40 knots, respectively, or a difference of 30 knots (15 m./sec.) across the front. In the Ft. Smith sounding, plotted in figure 8 of [2] the SSW wind increases from 20 knots below the front to 40 knots above the front, a difference of 20 knots or 10 m./sec. Thus, a value of $\Delta u_1=10$ to 15 m./sec. appears reasonable. From the Ft. Smith and Shreveport soundings 24 hours later (figure 16 of [2]) a reasonable value for Δu_2 appears to be 35 knots or 17.5 m./sec. The values of the frontal displacement in table 1 have been computed using the various values of the temperature and wind speed differences denoted above.

The observed value of 260 km. is larger than the values in the first column but smaller than the value computed for $\Delta u_1=\Delta u_2=10$ m./sec.

Turning now to the frontal motions along the 100th meridian, as depicted in figure 2, there are irregular oscillations—varying in period from 6 hours to 18 hours and in total amplitude from 90 km. to 140 km. The San Angelo, Tex., sounding (figure 3 of [2]) shows a cooling in the first kilometer of 3° to 5° from 0300 to 1500 GMT on May 11 and a spread of wet-bulb temperature depression from 10° to 17° C. In the last two columns of table 1, frontal displacements have been computed for $\Delta T_2=4°$C. first under the assumption that the frontal wind difference increases from 10 m./sec. to 17.5 m./sec., and second, that it is constant at 10 m./sec. For the first assumption $\delta=60$ km. and for the second, $\delta=140$ km.; the observed displacements are between these values.

TABLE 1.—*Computed frontal displacements ($H=2$ km., $T=290°$ K., $f=0.73\cdot10^{-4}$ sec.$^{-1}$ (30° lat.)*

ΔT_1 (° C.)	1	1	1	1	1	1
ΔT_2	7	7	7	7	4	4
Δu_1 (m./sec.)	10	15	10	15	10	10
Δu_2	17.5	17.5	10	15	17.5	10
δ (km.)	140	155	280	185	60	140

Thus, in summary, it may be said that if one interprets the oscillations of the fronts as precipitation-induced, then there is reasonably good agreement between computed and observed maximum displacements; the character and period of the frontal motions suggest inertial oscillations, sometimes damped by friction, and other times reintensified apparently by new surges of the rain-cooled air north of the front.

LIBERATION OF KINETIC ENERGY

To obtain an estimate of the kinetic energy liberated by the adjustment of the front, let us now compute the difference ΔE, between initial and final potential energy of the mass distribution shown in figure 1 *after the cooling has taken place.* Referring to figure 1 and making use of equation (2), we find for a 1-cm. thick vertical slab of air

$$\Delta E=\Delta E_c+\Delta E_w=\frac{gH^2}{6}(\rho'_2-\rho_1)\delta \tag{4}$$

Neglecting friction,

$$\Delta E=\frac{1}{2}M(\overline{V}^2-\overline{V}_0^2)=\frac{H^2}{4}\cot\alpha_2(\rho'_2+\rho_1)(\overline{V}^2-\overline{V}_0^2) \tag{5}$$

where M is the total mass of the 1-cm. thick air slab above line DE and $\overline{V}$, $\overline{V}_0$ are final and initial average wind speeds in the same mass.

Equating (4) and (5), and making use of equations (1) and (2),

$$\overline{V}^2-\overline{V}_0^2=\frac{gH}{3}\frac{\rho'_2-\rho_1}{\rho'_2+\rho_1}\left[1-\left(\frac{\rho_2-\rho_1}{\rho'_2-\rho_1}\right)\frac{\rho'_2\,\Delta u_2}{\rho_2\,\Delta u_1}\right]$$

$$=\frac{gH}{3}\frac{T_1-T'_2}{T_1+T'_2}\left[1-\left(\frac{T_1-T_2}{T_1-T'_2}\right)\frac{\Delta u_2}{\Delta u_1}\right]$$

Taking $H=2\cdot10^5$ cm., $\Delta u_2=\Delta u_1$, $T_1-T'_2=8°$, $T_1-T_2=1°$, $\frac{T_1+T'_2}{2}=290°K.$, and $\overline{V}_0=10$ m./sec.$=20$ knots, we find $\overline{V}=13.4$ m./sec.$=27$ knots.

Thus, 3.4 m./sec. or 7 knots is the average increase in wind speed in the system caused by flattening out of the front by precipitation cooling. This is an overestimate since friction is neglected. If the liberated kinetic energy is not spread over the entire system of air masses but is localized in certain areas this would mean high local winds, such as occurred in portions of the cool air mass near the front.

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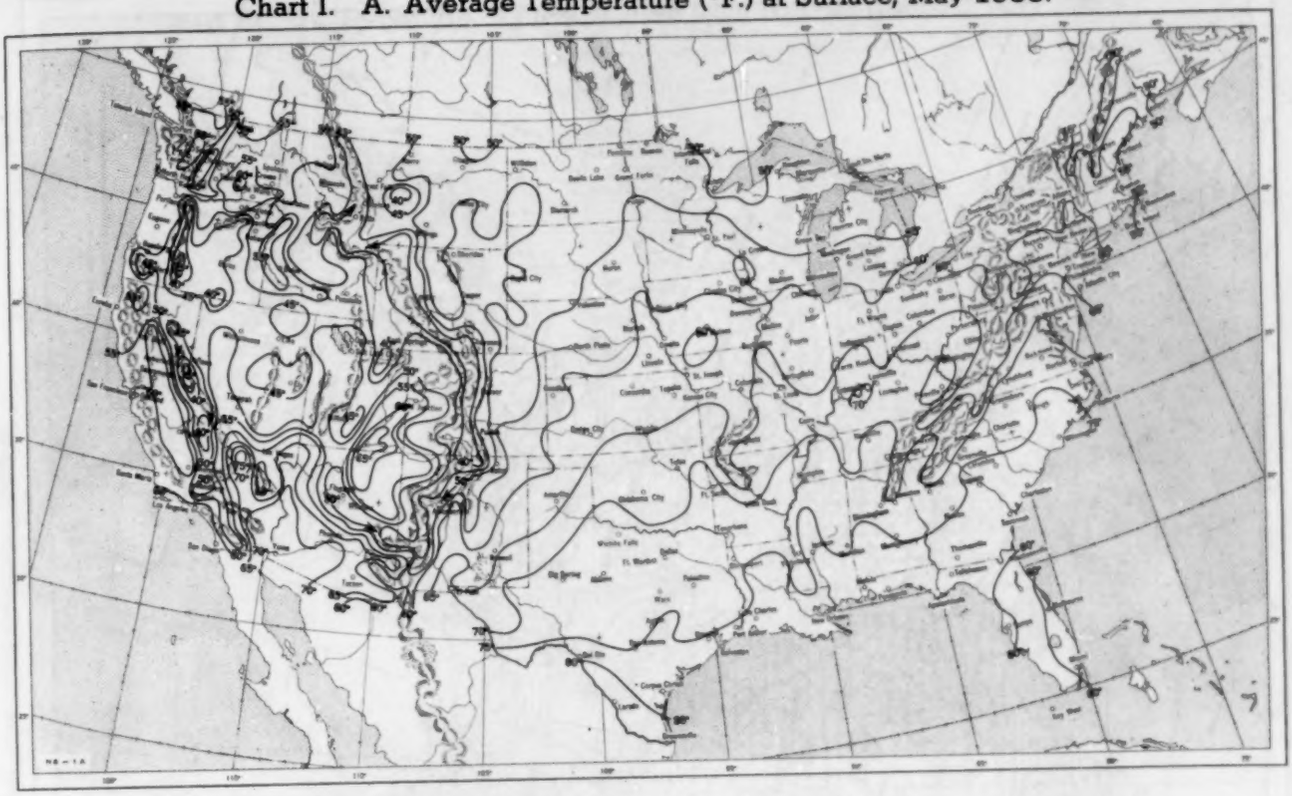
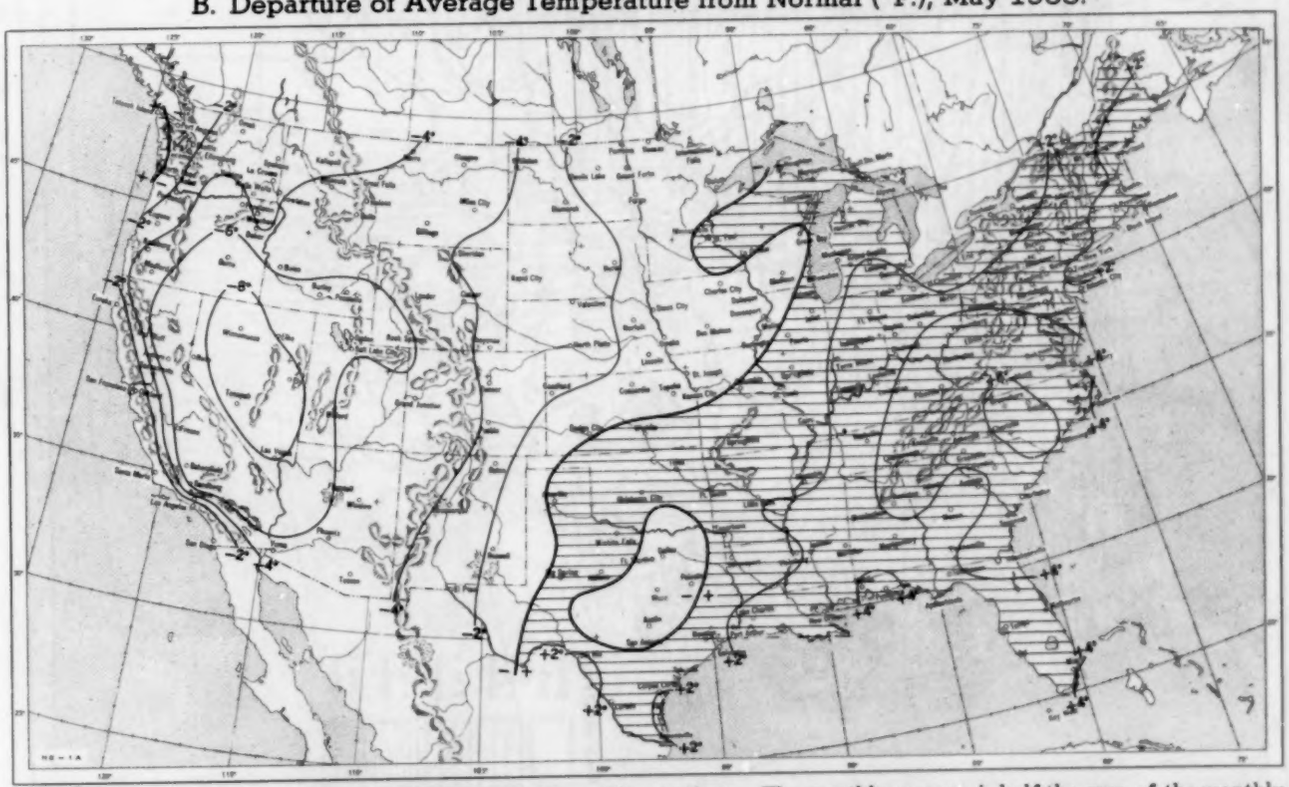
U. S. GOVERNMENT PRINTING OFFICE: 1953

Chart I. A. Average Temperature (°F.) at Surface, May 1953.



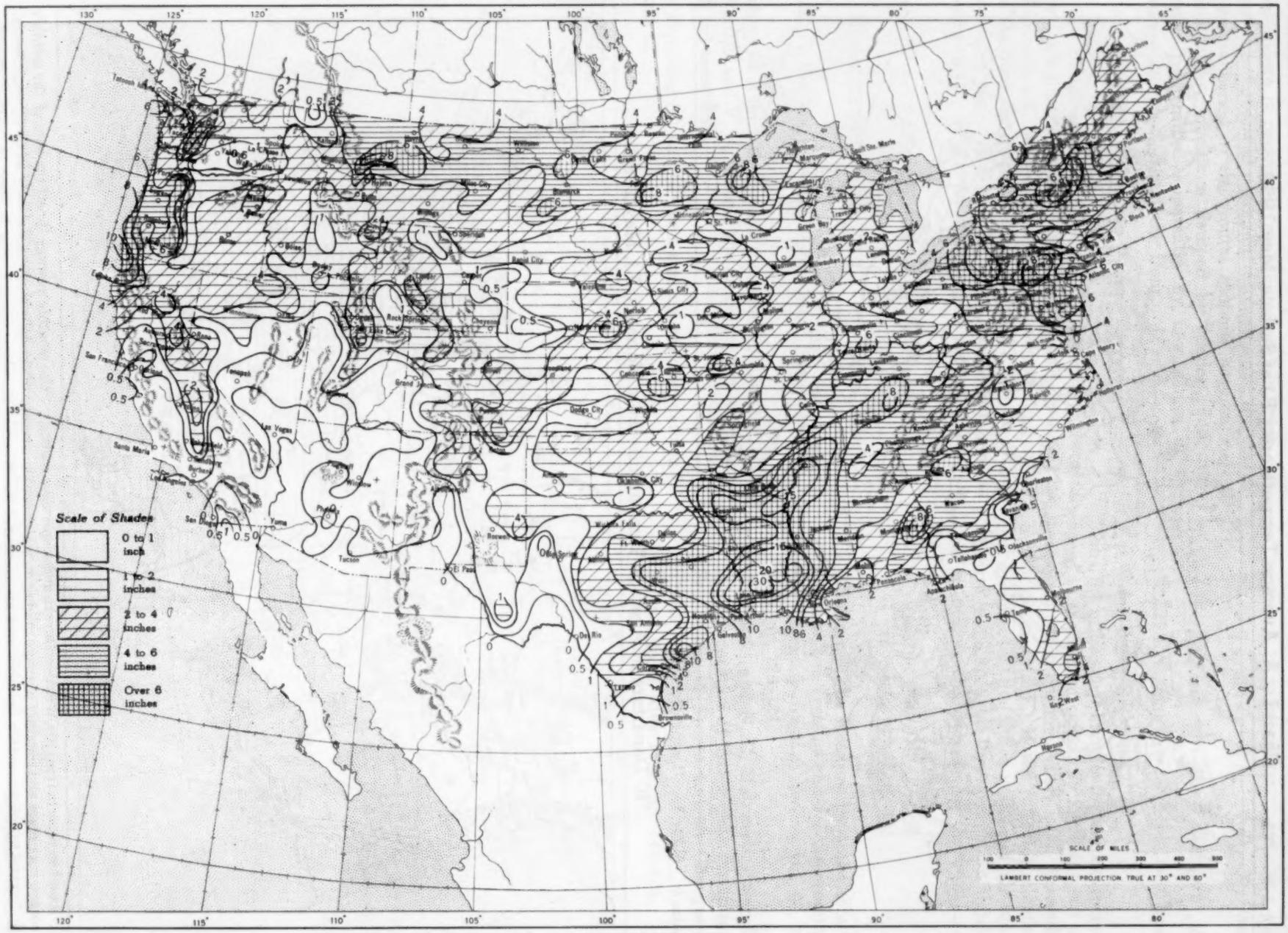
B. Departure of Average Temperature from Normal (°F.), May 1953.



A. Based on reports from 800 Weather Bureau and cooperative stations. The monthly average is half the sum of the monthly average maximum and monthly average minimum, which are the average of the daily maxima and daily minima, respectively.

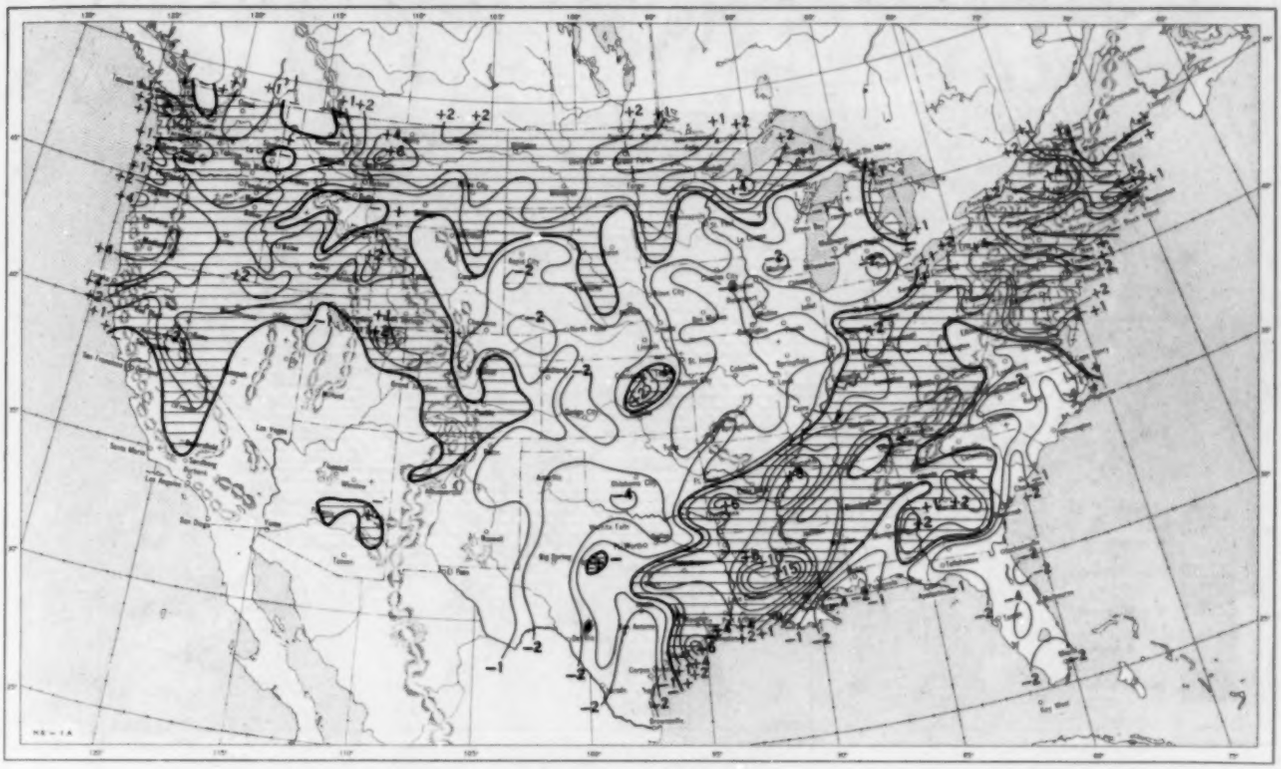
B. Normal average monthly temperatures are computed for Weather Bureau stations having at least 10 years of record.

Chart II. Total Precipitation (Inches), May 1953.

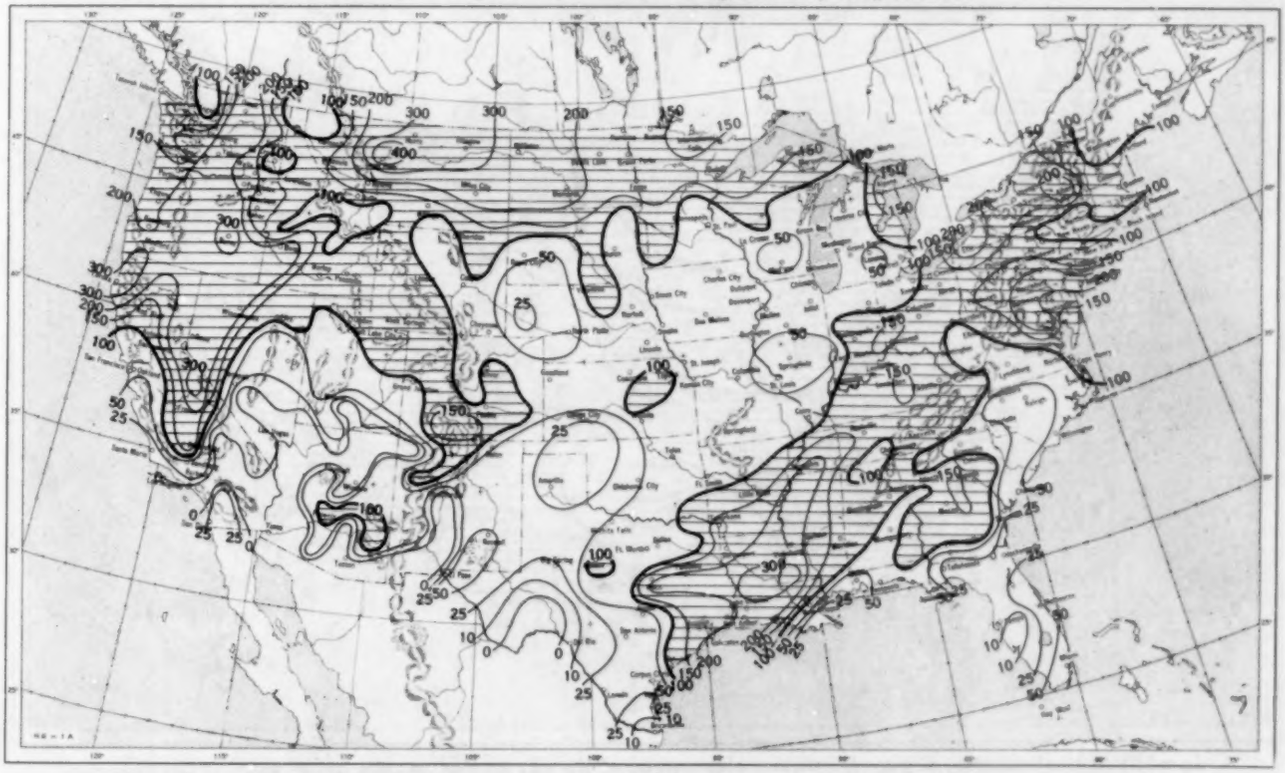

Scale of Shades
0 to 1 inch
1 to 2 inches
2 to 4 inches
4 to 6 inches
Over 6 inches
SCALE OF MILES
LAMBERT CONFORMAL PROJECTION TRUE AT 30° AND 60°

Based on daily precipitation records at 800 Weather Bureau and cooperative stations.

Chart III. A. Departure of Precipitation from Normal (Inches), May 1953.

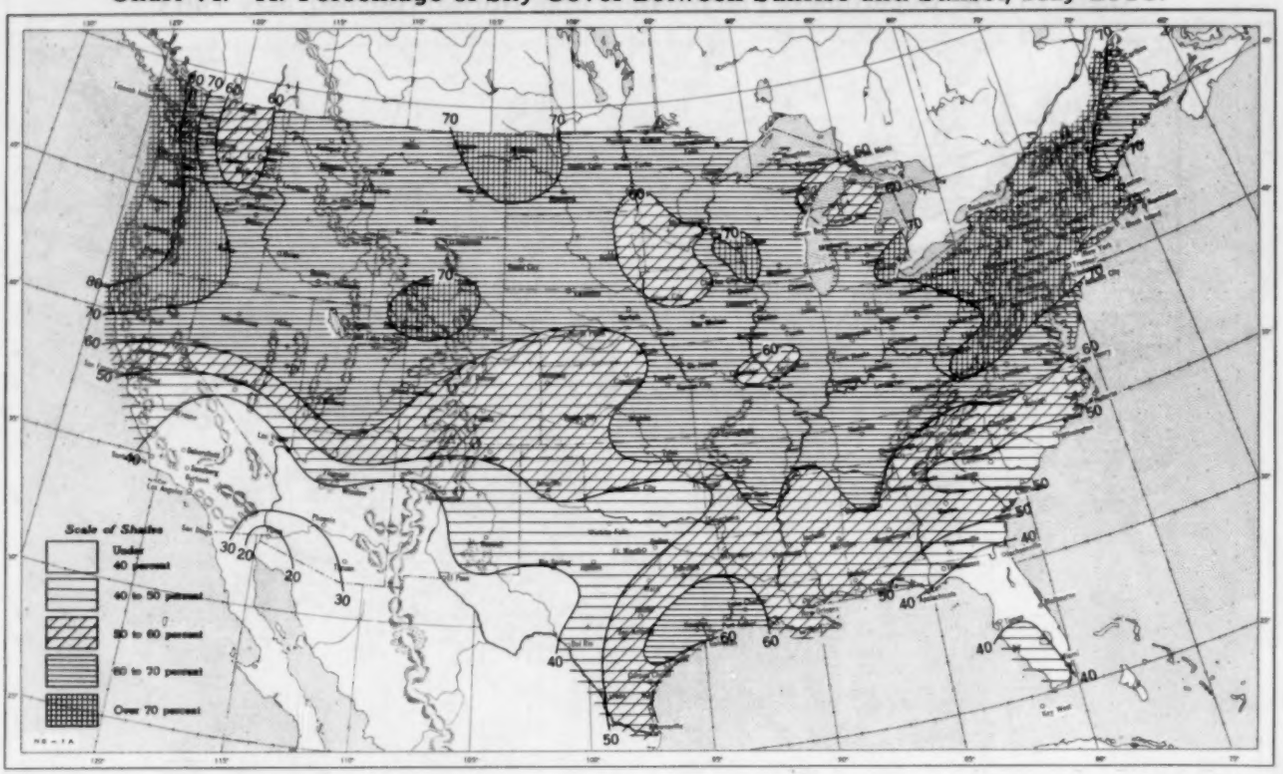


B. Percentage of Normal Precipitation, May 1953.

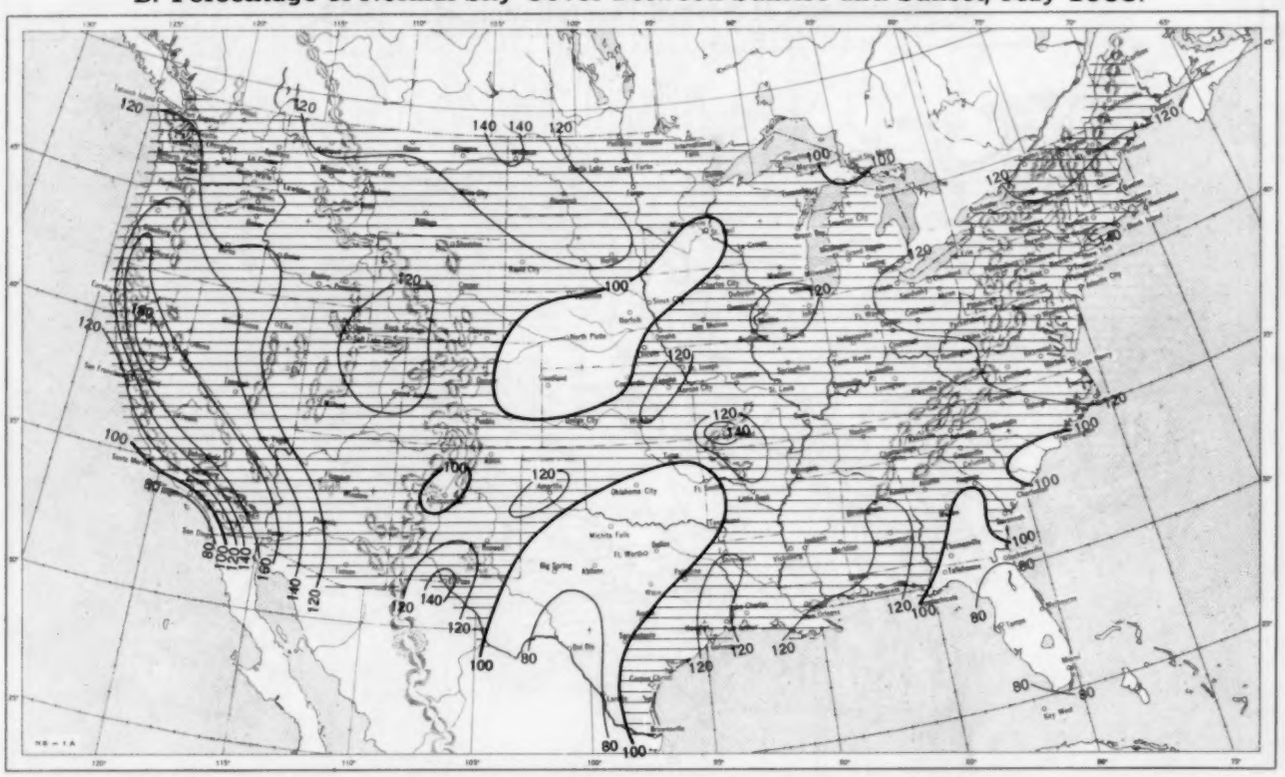


Normal monthly precipitation amounts are computed for stations having at least 10 years of record.

Chart VI. A. Percentage of Sky Cover Between Sunrise and Sunset, May 1953.

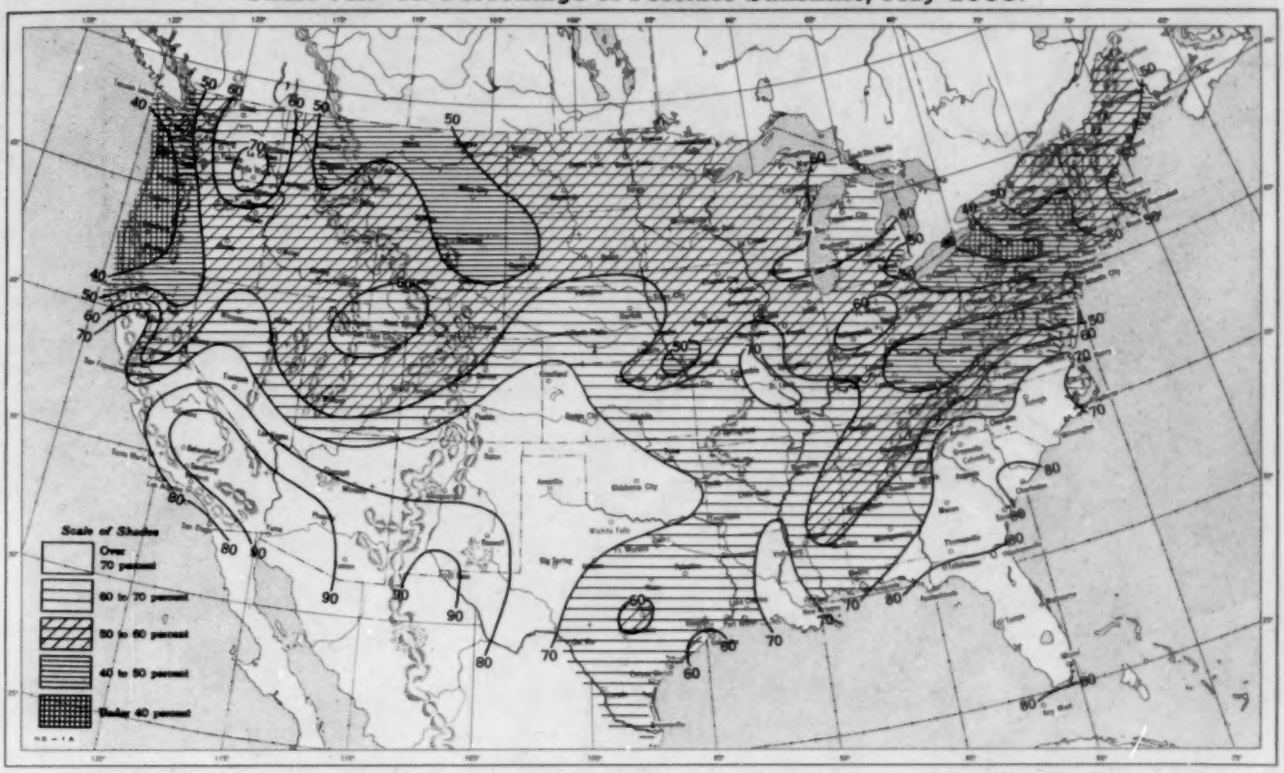


B. Percentage of Normal Sky Cover Between Sunrise and Sunset, May 1953.

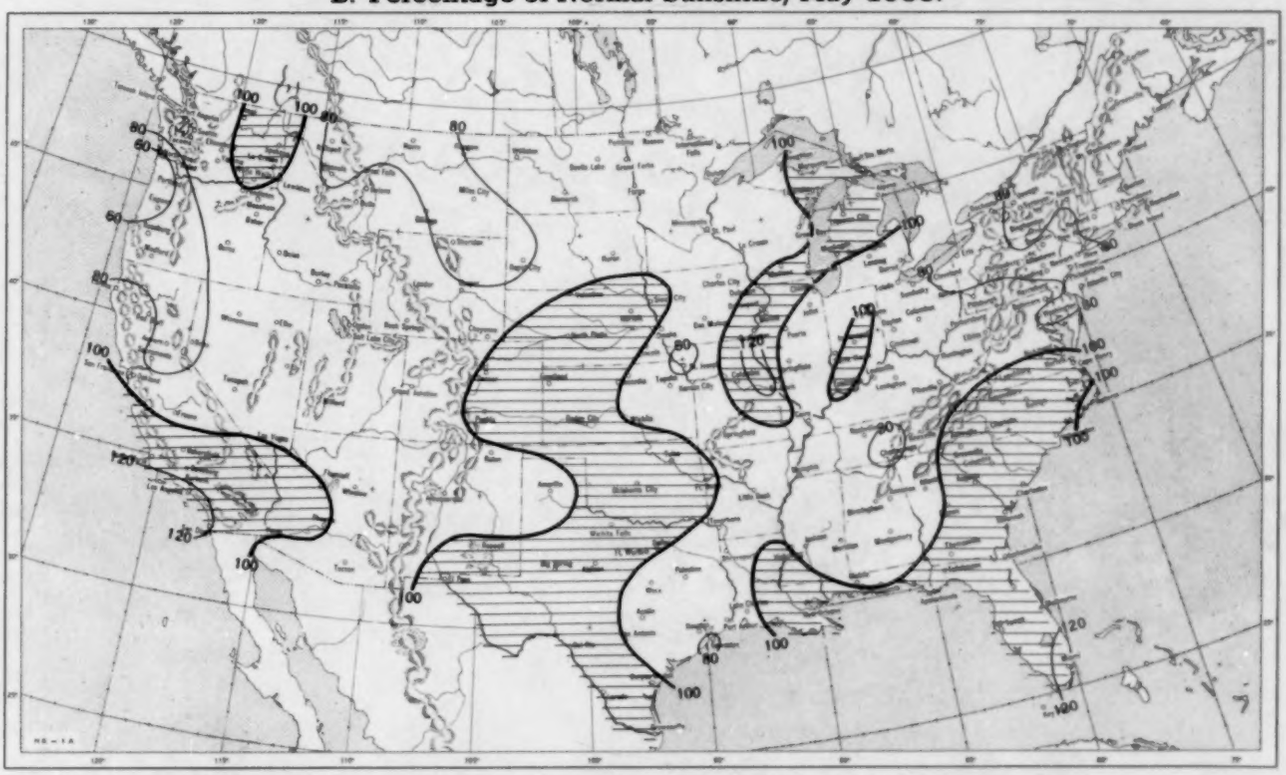


A. In addition to cloudiness, sky cover includes obscuration of the sky by fog, smoke, snow, etc. Chart based on visual observations made hourly at Weather Bureau stations and averaged over the month. B. Computations of normal amount of sky cover are made for stations having at least 10 years of record.

Chart VII. A. Percentage of Possible Sunshine, May 1953.



B. Percentage of Normal Sunshine, May 1953.



A. Computed from total number of hours of observed sunshine in relation to total number of possible hours of sunshine during month. B. Normals are computed for stations having at least 10 years of record.

Chart VIII. Average Daily Values of Solar Radiation, Direct + Diffuse, May 1953. Inset: Percentage of Normal

Average Daily Solar Radiation, May 1953.

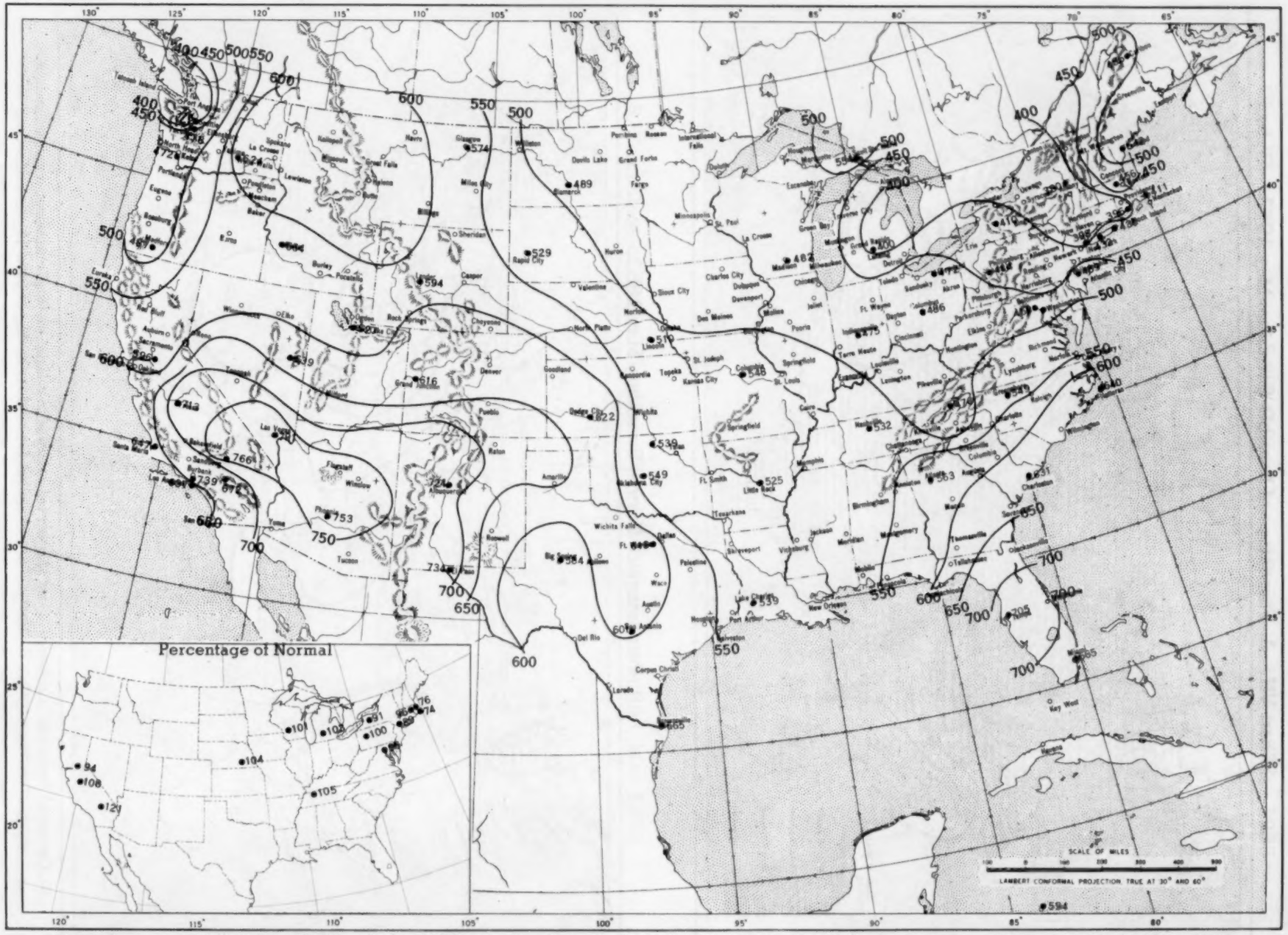
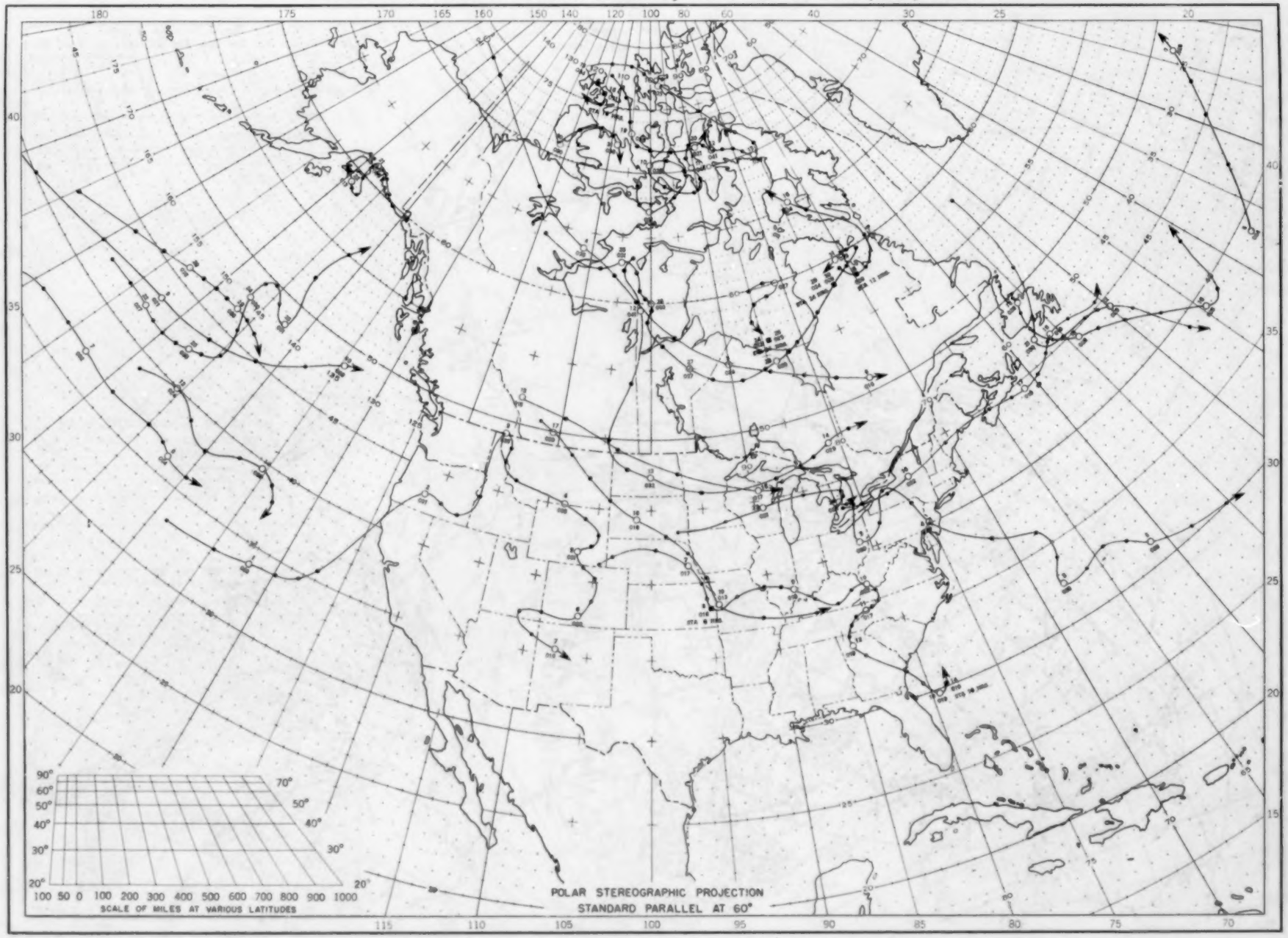


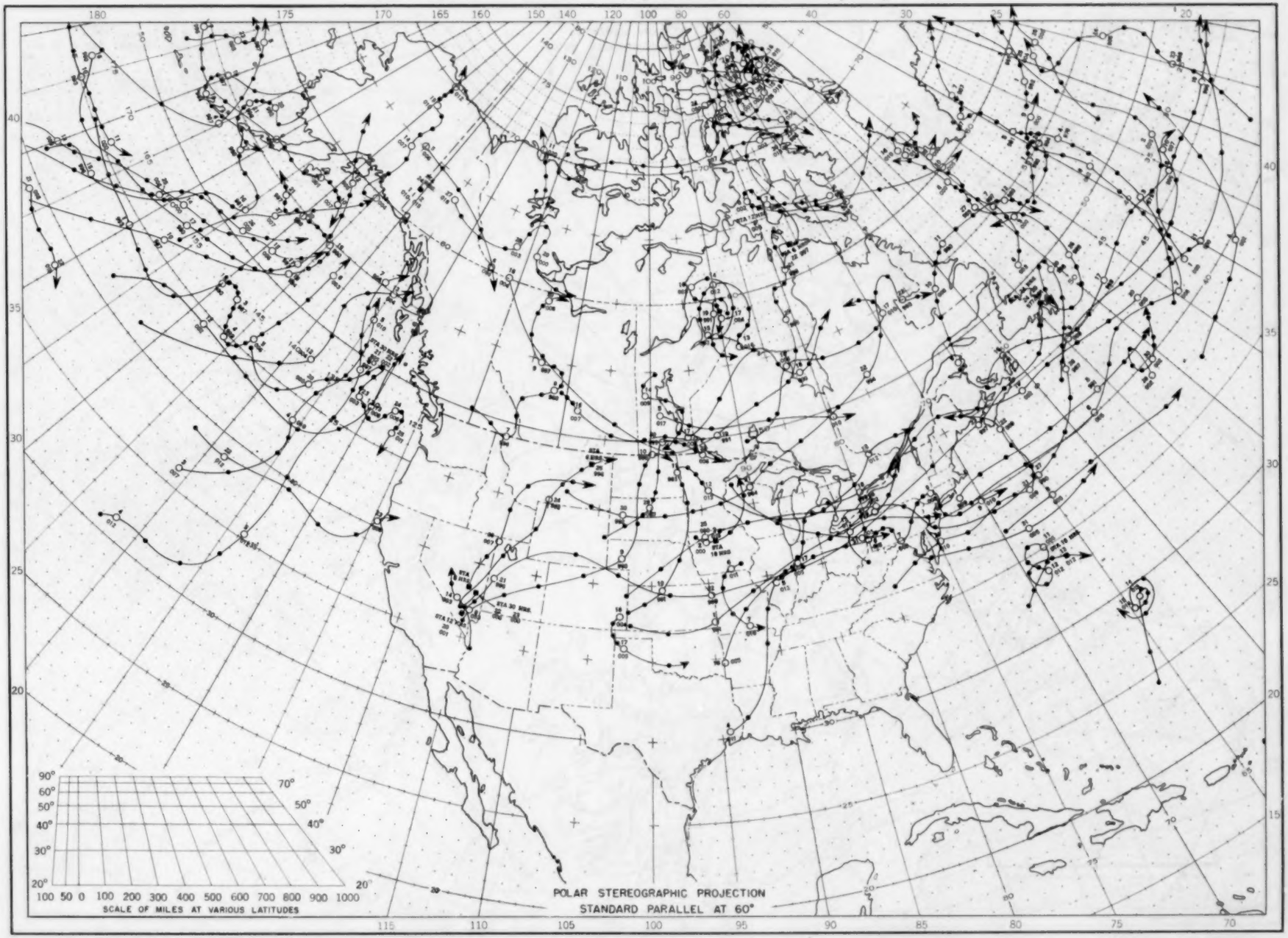
Chart shows mean daily solar radiation, direct + diffuse, received on a horizontal surface in langleys (1 langley = 1 gm. cal. cm. $^{-2}$). Basic data for isolines are shown on chart. Further estimates are obtained from supplementary data for which limits of accuracy are wider than for those data shown. Normals are computed for stations having at least 9 years of record.

Chart IX. Tracks of Centers of Anticyclones at Sea Level, May 1953.


POLAR STEREOGRAPHIC PROJECTION
STANDARD PARALLEL AT 60°
SCALE OF MILES AT VARIOUS LATITUDES
100 50 0 100 200 300 400 500 600 700 800 900 1000

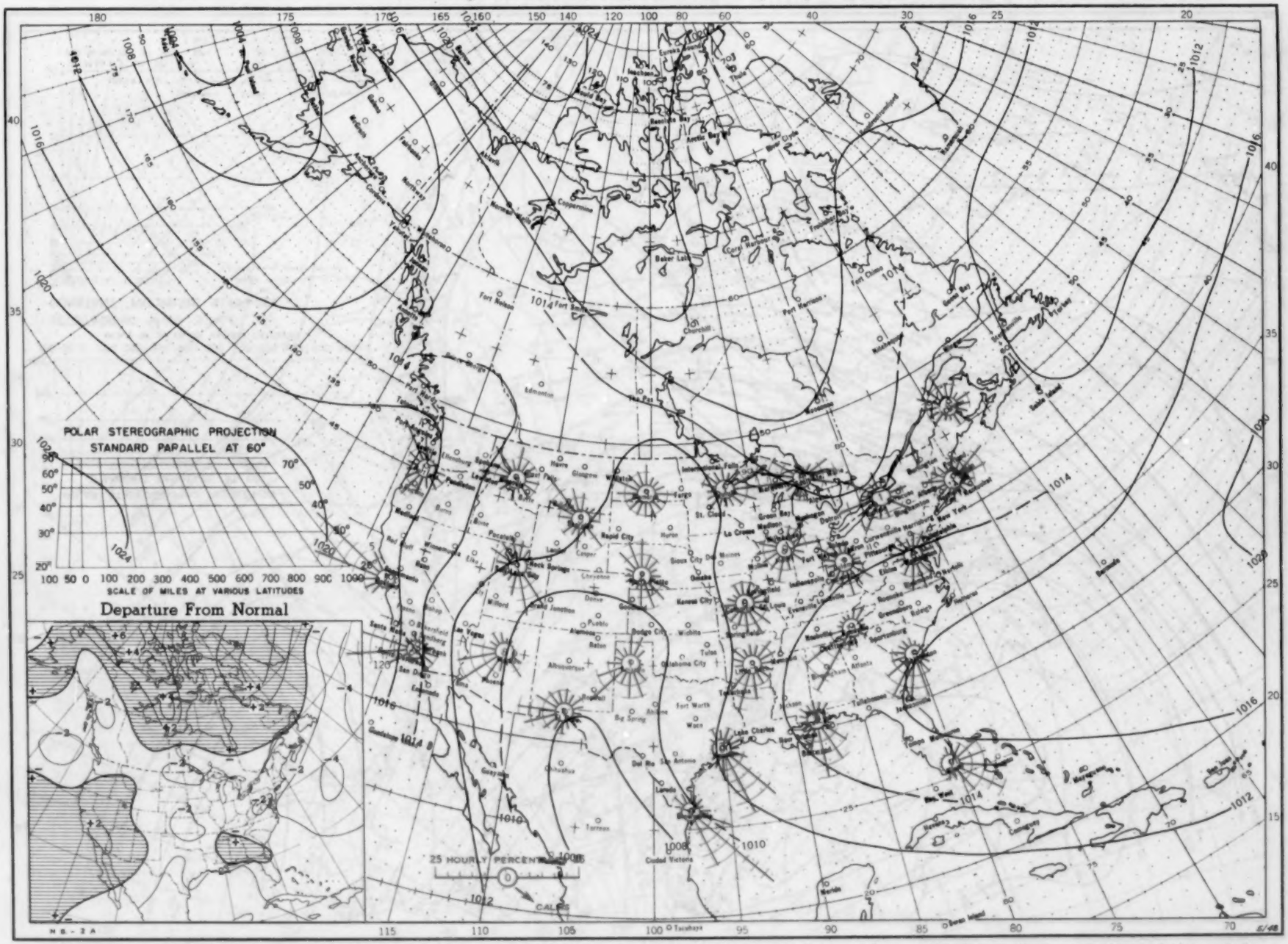
Circle indicates position of center at 7:30 a. m. E. S. T. Figure above circle indicates date, figure below, pressure to nearest millibar. Dots indicate intervening 6-hourly positions. Squares indicate position of stationary center for period shown. Dashed line in track indicates reformation at new position. Only those centers which could be identified for 24 hours or more are included.

Chart X. Tracks of Centers of Cyclones at Sea Level, May 1953.


POLAR STEREOGRAPHIC PROJECTION
STANDARD PARALLEL AT 60°
SCALE OF MILES AT VARIOUS LATITUDES
100 50 0 100 200 300 400 500 600 700 800 900 1000

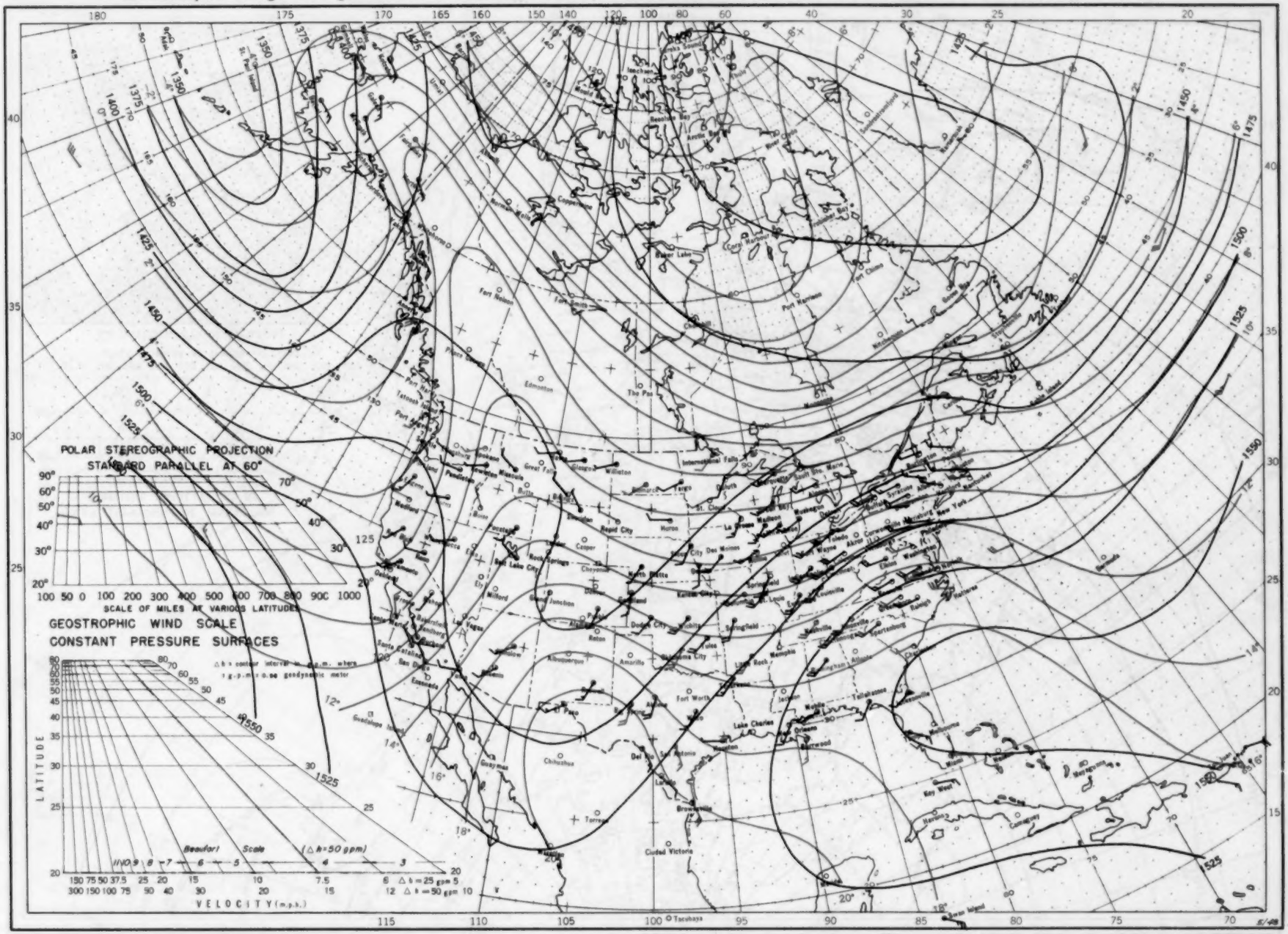
Circle indicates position of center at 7:30 a. m. E. S. T. See Chart IX for explanation of symbols.

Chart XI. Average Sea Level Pressure (mb.) and Surface Windroses, May 1953. Inset: Departure of Average Pressure (mb.) from Normal, May 1953.



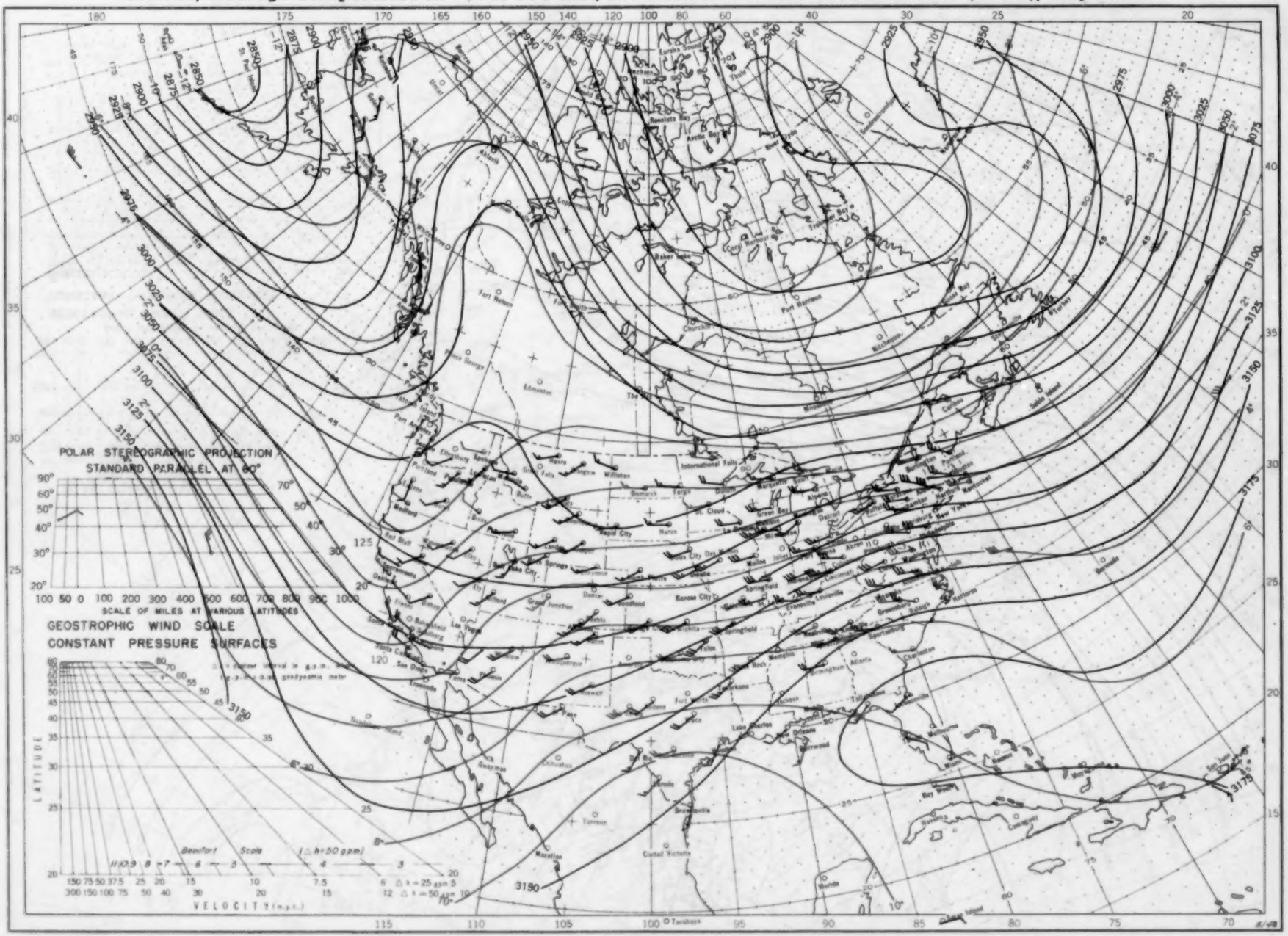
Average sea level pressures are obtained from the averages of the 7:30 a. m. and 7:30 p. m. E. S. T. readings. Windroses show percentage of time wind blew from 16 compass points or was calm during the month. Pressure normals are computed for stations having at least 10 years of record and for 10° intersections in a diamond grid based on readings from the Historical Weather Maps (1899-1939) for the 20 years of most complete data coverage prior to 1940.

Chart XII. Average Dynamic Height in Geopotential Meters (1 g.p.m. = 0.98 dynamic meters) of the 850-mb. Pressure Surface, Average Temperature in °C. at 850 mb., and Resultant Winds at 1500 Meters (m.s.l.), May 1953.



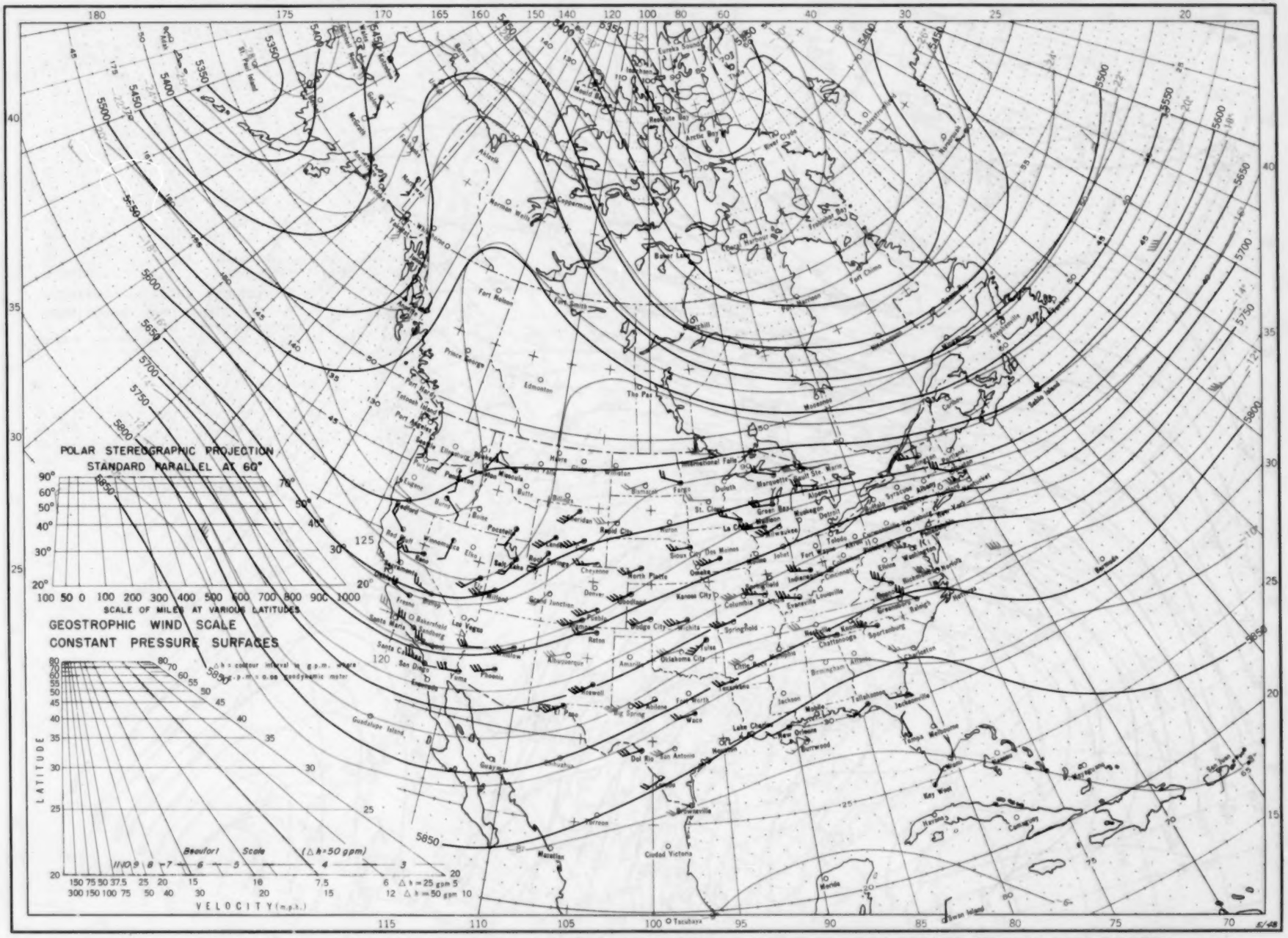
Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins taken at 0300 G. M. T.

Chart XIII. Average Dynamic Height in Geopotential Meters (1 g.p.m.. = 0.98 dynamic meters) of the 700-mb. Pressure Surface, Average Temperature in °C. at 700 mb., and Resultant Winds at 3000 Meters (m.s.l.), May 1953.



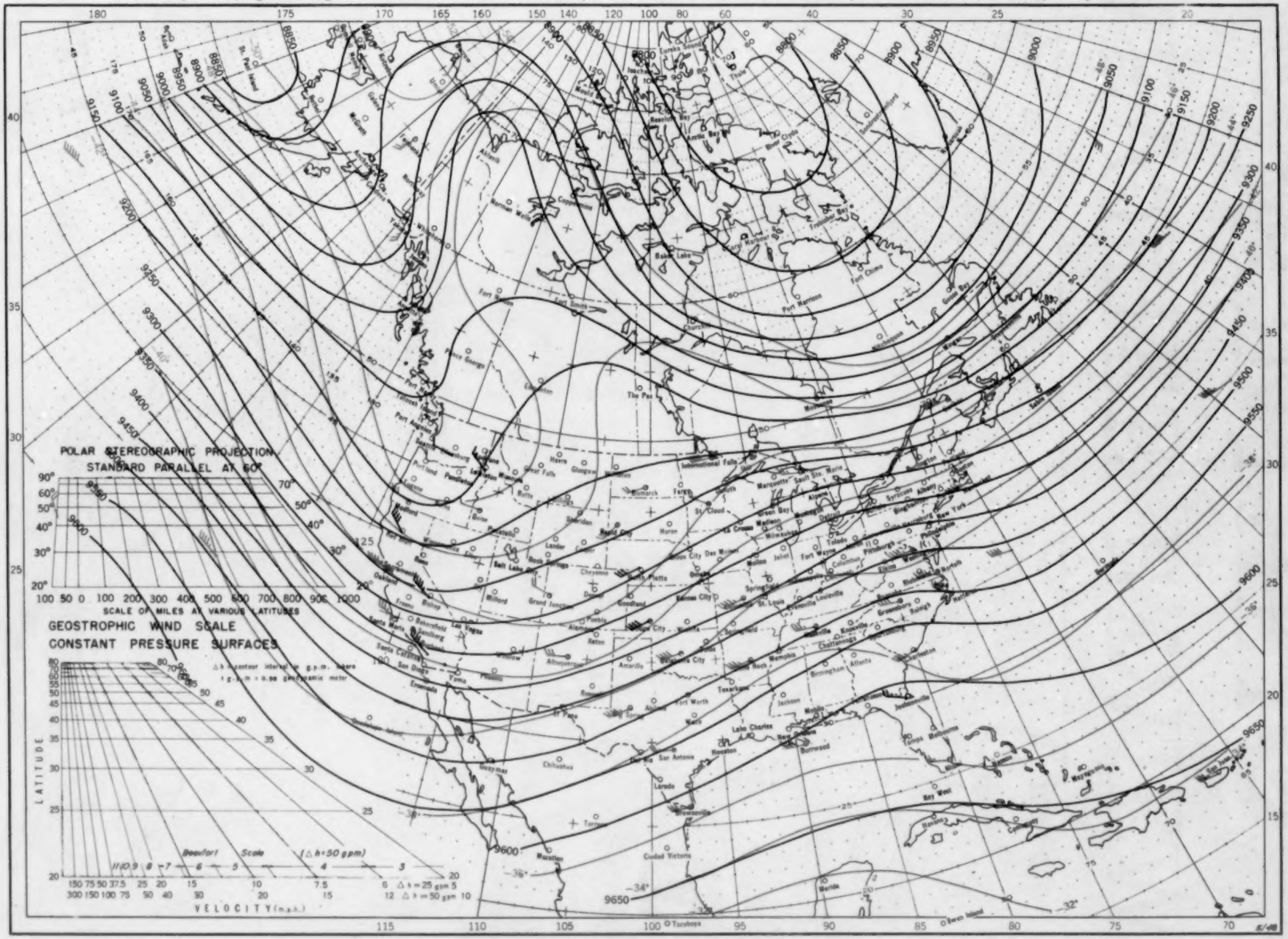
Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins taken at 0300 G. M. T.

Chart XIV. Average Dynamic Height in Geopotential Meters (1 g.p.m. = 0.98 dynamic meters) of the 500-mb. Pressure Surface, Average Temperature in °C. at 500 mb., and Resultant Winds at 5000 Meters (m.s.l.), May 1953.



Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins at 0300 G. M. T.

Chart XV. Average Dynamic Height in Geopotential Meters (1 g.p.m. = 0.98 dynamic meters) of the 300-mb. Pressure Surface, Average Temperature in °C. at 300 mb., and Resultant Winds at 10,000 Meters (m.s.l.), May 1953.



Contour lines and isotherms based on radiosonde observations at 0300 G. M. T. Winds shown in black are based on pilot balloon observations at 2100 G. M. T.; those shown in red are based on rawins at 0300 G. M. T.